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Lamont-Doherty Geological Observatory of Columbia University
Palisades, New York 10964

A COMPREHENSIVE STUDY OF THE SEISMOTECTONICS OF THE EASTERN
ALEUTIAN ARC AND ASSOCIATED VOLCANIC SYSTEMS

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By

K.H. Jacob, J.N. Davies, L.R. Sykes, L. House, J. Mori, J. Krause,
J. Boatwright, S. McNutt, J. Hauptman; J. Peterson, O.J. Perez
R. Bilham, J. Beavan, and M. Winslow

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ABSTRACT

The modern and historic seismicity of the eastern Aleutian arc has been thoroughly researched and shows clear evidence that the Shumagin Islands region of the arc: (a) has been ruptured at least partly by previous large or great earthquakes; (b) has recurrence times for large earthquakes between 50 and 115 years; (c) has presently the properties of a seismic gap, i.e. was not ruptured by a large or great earthquake for at least 76 years; (d) shows a 'donut' configuration of modest seismicity including high-stress-drop events around the rim of the expected rupture zone of a future large earthquake. From this combined evidence we conclude that the likelihood for a large or great earthquake to occur within the Shumagin Island gap is high (about 50% probability) within the next decade and very high (about 80% to 90% probability) within the next three decades. This study includes detailed determinations of stress, stress-drop and other source parameters of moderate earthquakes in the Shumagin Islands gap, and similar results for the St. Elias earthquake near the Yakataga gap in the eastern Gulf of Alaska. Volcano seismicity, crustal deformation and P-residuals continue to be monitored and analyzed in the eastern Aleutian arc segment.

INTRODUCTION

The primary results of research and field activities for the seismo-tectonic study of the eastern Aleutian arc, Alaska, carried out by Lamont-Doherty Geological Observatory during the Contract period March 1, 1979 to February 28, 1980 are summarized in this report. The report describes completed research for the first three quarters and outlines work in progress for the final quarter of this period.

This project is aimed to provide both a broad overview of the seismicity and tectonics of a typical active subduction zone and associated volcanic arc as well as a detailed in-depth study of the microearthquake-activity and crustal deformation in this arc and of its associated volcanic systems. For the in-depth study we operate a regional telemetered network of some 30 seismic stations covering a 600 km long arc segment; this network includes a subarray of twelve stations densely spaced on and around Pavlof Volcano on the western portion of the Alaska Peninsula for monitoring the seismicity associated with the magmatic processes in and beneath one of the most active volcanos of the Aleutian arc. The Pavlof volcano study is aimed towards basic volcanic research which at a later time will allow to realistically evaluate the geothermal potential of some Aleutian volcanos.

The cost for operating the eastern Aleutian seismic network is shared between this DOE-supported research program and a closely coordinated study under the Alaska Outer Continental Shelf Environmental Assessment Program (OCSEAP) administered by NOAA. OCSEAP urgently needs seismic hazards information for development of hydrocarbon resources on the vast

Alaska shelf regions. The difficult logistic and climatic conditions and the vast expanse even of this small segment of the Aleutian arc (600 km long) require a level of effort both in manpower, and logistic and financial support which would by far exceed the fiscal limitations set for either of the two complimentary programs. It is therefore fortunate that both projects rely on the same seismological data base and hence can share the high cost for maintenance and technical realization.

To keep in line with the goal of providing a broad seismic and tectonic overview over the eastern Aleutian arc we have devoted a good portion of our research efforts during this year to summarize both the historic and modern instrumental seismicity data. We have done this with emphasis on two particularly interesting segments of the Pacific/North American plate boundary in Alaska; both of these regions are "seismic gaps" in which large or great earthquakes are expected to occur in the next decade or two: i.e. the region of the Shumagin Islands seismic gap in the eastern Aleutians, and farther to the east, the Yakataga seismic gap in the eastern Gulf of Alaska. Excerpts from these completed (Davies and House, 1979; McCann et al., 1979; Perez and Jacob, 1980] or nearly completed studies (Davies et al., 1980) are included in this report.

In contrast to the broadly conceived seismotectonic studies in these two seismic gaps, we have carried out detailed analysis of particular earthquakes in both regions. The $M_s = 7.7$ St. Elias earthquake of February 28, 1979 which occurred in the eastern Gulf of Alaska provided a good opportunity to analyze teleseismic records (Boatwright, 1980) in a similar fashion as those for the 1974 earthquakes ($m_b = 5.8$ and 6) in the Shumagin Islands (House and Boatwright, 1980). The analysis yields source parameters such as stress drop, ambient stress, moment, and average dis-

placement. For the Shumagin Islands region a similar effort is in progress (J. Mori, unpublished data) using mostly teleseismic recordings from events for which new fault-plane solutions have been obtained (House and Jacob, 1980). The stress patterns, which emerge from these studies indicate that asperities of high-stress (~ 1 kbar) break along the rim of expected future rupture zones of great shallow thrust earthquakes. Moreover it appears that normal faulting events near trenches relieve lower stress drops (≥ 100 bar) than events on the shallow thrust zones. Stress drops associated with intermediate-depth earthquakes within the descending slab appear to be the lowest in magnitudes of stresses (< 100 bar) compared to those of the former two types of shallow events.

Of course a major effort was placed on routinely analyzing the incoming seismic network data to obtain as quickly as possible and as complete as possible a catalogue of local and regional earthquake locations. We have almost resolved to work up the back log of locations from previous years and only a six-months period remains.

The volcano research with its aim towards evaluating the geothermal potential of island-arc and subduction-related volcanism has been furthered along two lines: A most interesting result was obtained (S. McNutt, 1979) for correlating minor seismicity during Pavlof Volcano eruptions with tidal stress. The importance of this result lies in the fact that the correlation with compressional and dilatational components of tidal stresses is different for the beginning and cessation of an eruption; this allows some inferences on the physical processes that take place within and beneath the volcanic pile during magma emplacement, probably above the feeding magma chamber. The search for a magma chamber beneath Pavlof Volcano is yet unresolved but by no means do we have evidence that it does not exist.

The other volcanic study with implications for subduction-zone volcanism is of a broad regional character and is part of the completed eastern Gulf of Alaska study (Perez and Jacob, 1980). We have shown that the large cluster of Quaternary and Recently active Wrangell-volcanoes must be related to subduction of Pacific ocean floor, despite the fact that a dipping Benioff zone of intermediate-depth earthquakes has heretofore not been proven to exist within the hypothesized descending slab of Pacific lithosphere. We further concluded that an extension of this subduction-related volcanism towards SW into Canada and SE Alaska may exist. We also draw attention to widespread alkaline basalt volcanism in SE Alaska and adjacent British Columbia by showing its relation to the extensional tectonic setting within this highly oblique subduction zone which traditionally has been interpreted as a pure transform plate boundary. A reevaluation of the geothermal potential of those entire regions in the light of these tectonic interpretations may be warranted. At least one of the potential sites (NE of Ketchikan, SE Alaska) has hot water systems with reported temperatures in excess of 150°C (U.S.G.S., Muffler (edit.), 1970).

Our crustal deformation studies in the Shumagin area continued through precise releveling of geodetic lines, sea-level meter operations (Bilham and Beavan, 1979) and terrace studies (Winslow, 1978). These efforts have been partially (and in the future will probably be completely) supported by other funding sources (NSF, USGS) since these projects have in size and effort outgrown the format of being a small program element within this DOE-supported research.

The technical field efforts have been concentrating on consolidating the quality and reliability of the seismic network and to develop and

deploy automatically editing tape-recording devices at our two central recording sites at Dutch Harbor (completed) and Sand Point (to be installed in remaining quarter). Unfortunately we still rely on analog data recording and hence must spend considerable efforts to transfer data into digital format for processing on the new USGS-provided computer specifically devoted to seismic network data analysis. This unfortunate fact stems from the passed fiscal limitations and we will seek remedy in forthcoming project periods.

The following sections contain more detailed information on those topics briefly summarized as well as from those program elements not yet discussed in this brief introduction.

A Review of the Seismic Potential of the Shumagin Islands Seismic Gap.

The following text is an expanded version of the conclusions from a paper by J.N. Davies, L.R. Sykes, L. House and K.H. Jacob (1980, in preparation), which summarizes our present understanding of the seismicity and tectonics of the seismic gap centered in the Shumagin Islands region in the eastern Aleutian arc. These conclusions are only insofar modified from the original text as to allow reference to the pertinent figures and diagrams (Figures 1 through 9). The major thrust of this paper is to reevaluate the seismic potential of this seismic gap for a forthcoming large earthquake in the light of latest data.

We have summarized the seismotectonic history of the eastern Aleutian arc, focusing on the question of the existence of a seismic gap in the Shumagin Islands area. This summary includes: (1) a re-evaluation and extension of the record of great, thrust-type earthquakes (Figure 2) and

their aftershock zones that suggests a seismic gap with high potential for a great earthquake exists in the Shumagin Islands area (Figure 1); (2) a computation, based on both the historical record and theoretical considerations, of the expected repeat-times for earthquakes within and near this gap that suggests a high probability for a great earthquake within about a decade; (3) a consideration of the geologic and tectonic history (Figure 3) of this region with particular attention to major structural features that suggests that important transverse tectonic boundaries coincide with the southwest end of the aftershock zone of the great Alaskan earthquake of 1964 and with the eastern end of the aftershock zone of the Andreanof-Fox Islands earthquake of 1957; (4) a presentation of a seismicity map (Figure 4) and hypocentral cross-sections (Figures 5 and 6) based on data from the local, L-DGO operated, seismic network that show a near-absence of even low-level activity along the plate-plate interface, or main-thrust zone (MTZ), coupled with a very high level of activity at the down-dip end of the MTZ. These features suggest the MTZ is "locked", and further imply that strain energy is accumulating in preparation for a large earthquake, and that high stress levels already exist at least at the lower end of the MTZ; (5) presentation of new seismicity maps (Figures 7 and 8), based on teleseismic observations, of the shallow (≤ 80 km) earthquake activity of the eastern Aleutian arc for the past three and one-half decades that show the formation around the Shumagin gap of a "Mogi-donut" within the past fifteen years (Figure 7); (6) a discussion of new focal mechanisms [House and Jacob, 1980] that shows (Figure 9) the Shumagin area is, in terms of its general pattern of tectonic deformation, identical to the adjacent regions of the Aleutian arc. We expect, therefore, as in those regions, that strain is accumulating from the subduction of the

Pacific plate and will be released in a great earthquake. Following are more detailed statements of these conclusions.

Aftershock zones of the 1938, 1946 and 1948 earthquakes. A re-examination of Sykes' [1971] questioned extension of the 1938 aftershock zone (Figure 2) west of about 159°W shows that it was based on only two earthquakes one of which is not clearly an aftershock and both of which occurred at depths too great to be along the plate-plate interface. A restricted aftershock zone length of about 330 km is also more consistent with rupture areas derived from realistic values for seismic moment and the locus of tsunami generation for the 1938 event.

The moment of the 1946 event is too small for the aftershock area to have extended much east of Sanak Island (162°W). We attribute the large tsunami magnitude of this event to its unusual location beneath the inner trench wall and an exceptionally efficient mechanism of tsunami generation from imbricate faulting of the low rigidity sediments comprising that wall.

Similarly, the moment of the 1948 event is far too small to have broken much of the area between the 1946 and restricted 1938 zones. This event is also too deep to have occurred on the plate-plate interface, so we interpret it as having occurred within the down-going slab.

We conclude, therefore, that the approximately 250 km long area between the 1946 and restricted 1938 zones is a seismic gap that has not broken since at least 1903; this is termed the Shumagin Gap.

Possible previous ruptures of the Shumagin Gap. For a seismic gap to be regarded as capable of generating a great earthquake McCann et al. [1979] specified that it should have a history of at least one such earth-

quake. For the Aleutians, in general, the historic record is quite short. In the eastern Aleutians, however, the Russian settlements between Unalaska and Kodiak Island provide us with at least a partial record. A re-examination of derivative sources and some new translations (Kisslinger, 1980) of original records of these settlements provided the early part of the following record of earthquakes (Figure 2) that might have broken the Shumagin Gap, or at least portions thereof:

- 1788 - clearly a great earthquake that ruptured from Kodiak to Unga Island (in the Shumagins) and probably to Sanak Island.
- 1792 - a large, possibly great earthquake, only report is from Kodiak Island.
- 1844 - a strong earthquake, only report is from Kodiak Island, possibly not shallow.
- 1847 - a probably great earthquake that ruptured at least from Chirikof Island (in the Semedis) to Unga Island in the Shumagins.
- 1848 - a strong earthquake, only report is from Chirikof Island (in the Semidis).
- 1854 - a large, possibly great earthquake, only report is from Kodiak Island.
- 1868 - a "slight" earthquake, only report is from Unga, in the Shumagin Islands, tsunami report of 7 m on Unga is probably vastly overstated.
- 1880 - a major earthquake, only report is from Chirikof in the Semidi Islands.

- 1899 - $M_s = 7.7$, uncertain instrumental location, felt reports from Unalaska Island and Unga in the Shumagin Islands.
- 1900 - $M_s = 8.1$, originally placed near Yakataga but most strongly felt and most aftershocks at Kodiak Island.
- 1903 - $M = 8.3$, $H = 100$ km(?), epicenter at the eastern end of the 1938 aftershock zone.

Of these only the 1788 and 1847 events seem likely to have broken at least half of the Shumagin Gap; it is possible that part or all of the gap also broke in the 1899, 1900 and 1903 sequence of events. While not unequivocal, this record is an adequate basis for the conclusion that the Shumagin Gap has broken in at least one previous great earthquake.

Repeat times for the Shumagin Gap. From the historical record, above, it appears that part or all of the Shumagin Gap has broken at least once and perhaps twice in the 111-115 years between 1788 and 1899-1903. Similarly, it appears that the 1938 zone has broken at least twice and perhaps three times in the 150 years between 1788 and 1938. These observations yield average repeat times between 50 and 115 years for this part of the Aleutian arc. From plate-tectonic considerations of the width of the main-thrust zone ($W = 100$ km) and the velocity of convergence ($v_c = 7.4$ cm/year) we calculate repeat times of 66 to 86 years for the Shumagin Gap. Prorating a mean repeat time of 6.3 years between great, shallow earthquakes for the entire (4100 km) Aleutian arc to the Shumagin Gap (250 km) yields a repeat time of about 103 years. We note that it has been at least 75 years since the last great earthquake (1903) that could have broken the Shumagin Gap and quite possibly much longer. We also note that Kelleher [1970], using a different technique (space-time progres-

sions) from any of the above concluded that a great earthquake was likely in the 1938 zone by 1980. We conclude that even by the most conservative estimate of repeat times, a great thrust-type earthquake is highly likely in the Shumagin Gap within the next three decades and that the probability for the imminent occurrence of this earthquake is also significantly high.

Geologic and Tectonic Evidence for Transverse Structures. From a review of the geologic and tectonic history of the eastern Aleutian arc two structural features stand out (Figure 3): one is an inferred boundary perpendicular to the trend of the arc that cuts across the Alaska Peninsula in the vicinity of Becharof Lake and Wide Bay and extends southeast at least to the shelf edge along the southwest end of Kodiak Island, and the second is a series of grabens that are subparallel to the edge of the Bering Shelf and cross the arc obliquely in the vicinity of Unimak Island. Both of these transverse features have geologic, tectonic, volcanic and seismic expression; in the latter case both in terms of presently active micro- and meso-seismicity and the boundaries of aftershock zones of recent great earthquakes.

The inferred boundary through Wide Bay appears to truncate several geologic features that extend along strike of the arc to the northeast; among them are the Alaska-Aleutian Range batholith of Jurassic age, the Bruin Bay Fault, the Permian-or-older terranes of southwest Alaska, and on Kodiak Island, the Uyak-McHugh complex, another belt of Jurassic plutons, and on the shelf the Kodiak-Montague Island Fault and associated anticlines. Tectonic expression is implied by the inferred extension of the Alaska-Aleutian Range Batholith beneath the Quaternary sediments to the west of Becharof Lake and by the abrupt change in the style of folding from long, broad, gently dipping anticlines and synclines to the northwest of

Wide Bay to the short, narrow, tightly folded anticline and synclines to the southwest. There is a change in trend of the volcano-line just southwest of Becharof Lake which, perhaps coincidentally, was the site of the Ukikrek Maars eruptions in April of 1977. The outer Kodiak shelf just south of the western end of the Island is the site of persistent micro- and meso-earthquakes. Finally the transverse boundary through Becharof Lake, Wide Bay and the southwesternmost Kodiak Islands is also almost exactly colinear with the southwest end of the aftershock zone of the great Alaska earthquake of 1964.

The oblique boundary subparallel to the edge of the Bering Shelf (Figure 3) which cuts across the arc at Unimak Island is defined by the trend of the St. George and Sanak grabens. Unimak Island lies between these grabens and is almost entirely covered by Quaternary volcanics and associated sediments. This is in sharp contrast to wide exposures of Tertiary sediments for great lengths along the arc in either direction from Unimak Island. This blanket of Quaternary volcanics is the result of the unusual concentration of five major volcanic centers in the space of about 100 km along Unimak Island. Minor seismic activity is associated with both the St. George and Sanak Grabens. Unimak Island is close to the site of the unusually tsunamigenic Aleutian earthquake of 1946 and near the end of the aftershock zone of the great Andreanof-Fox Islands earthquake of 1957. Finally, this boundary also marks the transition from the oceanic arc to the west to the continental arc to the east.

It seems likely that these two transverse boundaries which have such strong geologic and tectonic expression probably controlled the extent of the ruptures and hence the aftershock zones of the 1964 and 1957 earthquakes, respectively. It is probable that they similarly will control the extent of future great earthquakes in the central and eastern Aleutian arc.

Definition of the Main-Thrust Zone and associated crustal seismicity from local network data. Cross-sections (Figures 5 and 6) of hypocenters determined using data (Figure 4) from L-DGO's network in the Shumagin Islands reveal that the plate-plate interface above a depth of approximately 30 km is presently aseismic. We interpret this as the result of the overthrusting North American plate being temporarily "locked" onto the descending Pacific plate. On either side of this locked interface, or main-thrust zone, strain energy is accumulating that will be released in a great earthquake. The most dense cluster of hypocenters in the Shumagin Islands cross-section (Figure 5) occurs at the down-dip edge of the main-thrust zone. This cluster consists, for the most part, of persistent aftershocks of two earthquakes of about magnitude six that had very high stress-drops [600-900 bars, House and Boatwright, 1980]. The most concentrated seismic activity in the crust occurs in an inclined band that dips steeply, is almost perpendicular to the lower main-thrust zone, and "outcrops" along the southeastern shore of the Alaska Peninsula and adjacent inner shelf. We interpret the high-stress-drop earthquakes, the associated high level of aftershock activity and the concentrated crustal activity that is "rooted" in the lower main thrust zone as the result of stress concentrations along the lower edge of the main-thrust zone. This stress concentration is the consequence of the abrupt transition at the aseismic front from the locked condition of the two plates across the main-thrust zone to a state of very low or no coupling between the plates below a depth of about 40 km.

Teleseismic Seismicity Maps and the Formation of a "Mogi-Donut". Maps of teleseismic seismicity in the eastern Aleutians for the past few decades show that the area between the aftershock zones of the Andreanof-

Fox Islands earthquake of 1957 and the great Alaska earthquake of 1964 is characterized by a relatively low level of seismic activity in comparison to adjacent portions of the Aleutian arc. To examine the pattern of interplate seismic activity we present two maps (Figures 7 and 8) of teleseismically derived epicenters whose depths were reported as less than 80 km. The first map (Figure 7) shows epicenters for earthquakes of $M \geq 5$ which occurred during the period January 1965-June 1979 and the second (Figure 8) for January 1945-December 1964.

The principal conclusion drawn from these maps is that the main-thrust zone is almost aseismic from mid-Unalaska to approximately the midpoint between the Shumagin and Semidi Islands, with the notable exception of the tsunamigenic Aleutian earthquake of 1946 and its aftershock (seaward of Unimak Island). Most of the earthquakes that do occur within this Unalaska-Semidi area are at depths of 40-60 km, along the down-dip edge of the main-thrust zone. The 1946 events divide the Unalaska-Semidi area into the "Unalaska Gap" to the west and the "Shumagin Gap" to the east (Figure 2). The main differences that are noted on the more recent map (Figure 7) compared to the one covering the earlier period (Figure 8) are the cessation of activity in the trench to the west of the Unalaska Gap, the occurrence of activity in the trench along the Shumagin Gap, and a slight westward movement of the eastern boundary of the Shumagin Gap. Most of the activity trenchward and eastward of the Shumagin Gap has occurred within the past fifteen years. We interpret the pattern that now exists around the Shumagin Gap as a "Mogi-donut" or a torus of activity around the edges of the future rupture zone of a great earthquake. The pattern around the Unalaska Gap is less well defined and overlaps the aftershock zone of the 1957 earthquake, so its interpretation is less

clear. We conclude that the most likely rupture zone for the next great earthquake in the eastern Aleutians is that defined by the Hogi-donut around the main-thrust zone of the Shumagin Gap (extending from mid-Unimak to midway between the Shumagin and Semidi Islands). It is possible, however, that once nucleated in the Shumagin Gap such an earthquake could rupture either the Unalaska Gap or the area of the aftershock zone of the 1938 earthquake.

Focal Mechanisms for earthquakes in the eastern Aleutian arc.

Consideration of an essentially new set (Figure 9, Table 1) of focal mechanisms [House and Jacob, 1980] leads to almost identical conclusions to those drawn by previous investigators regarding the general tectonic pattern in the eastern Aleutians. This pattern consists of normal faulting in the trench, shallow angle underthrusting between the plates beneath the shelf and failure resulting from either down-dip tension or compression within the descending slab of Pacific lithosphere at depths greater than about 60 km. There are two unusual strike-slip mechanisms for events under the inner wall of the trench, but because of the lack of depth control for these events and the critical dependence of their interpretation upon depth we refrain from further comment on these events other than to note their existence. For the intermediate depth events those to the east of Unimak Island show down dip tension whereas those to the west show compression. There are several possible interpretations of this pattern: it could reflect the different tectonic environments, oceanic to the west and continental to the east; it could reflect the temporally differing stress regimes related to the existence of the Shumagin Gap; or it could simply be a statistical aberration in our small sample. At present we cannot choose among these options. Since the general tectonic pattern as revealed by

focal mechanisms within the Shumagin Gap is identical to that of adjacent portions of the Aleutian arc we conclude that, like these adjacent portions, as the Pacific plate is underthrust beneath it, the Shumagin Gap region accumulates strain energy that is periodically released seismically rather than aseismically.

Summary. On the basis of a re-evaluation of the extent of the aftershock zone of the 1938 Aleutian earthquake we identify the Shumagin Seismic Gap between the restricted western end of the 1938 aftershock zone and the eastern end of the 1946 aftershock zone. This identification is corroborated by the existence of a Mogi-donut around the gap, the lack of seismic activity on the main thrust zone and the occurrence of high-stress-drop earthquakes at the base of the main thrust zone. The history of previous large earthquakes in the eastern Aleutians was extended by original translations of records made by early Russian settlers and explorers. This history suggests that the region of the Shumagin Gap has broken in at least one great earthquake in 1788 and possibly others in 1847 and 1898-1903. This fact plus the observation of the general tectonic similarity of this to adjacent regions that have recently broken in great earthquakes implies that the Shumagin Gap is not the result of aseismic slip, but that it releases plate motion seismically and hence can be expected to produce a large earthquake. Several lines of evidence lead to estimates of repeat times for the Shumagin Gap of 50 to 115 years. It has been 75 years since the last possible rupture of the Shumagin Gap. A high probability exists, therefore, for a great earthquake within one to three decades in the Shumagin Gap. The probability for the imminent occurrence of this event is also high. It is possible that when this earthquake occurs it will also rupture either the Unalaska Gap to the west or the 1938

aftershock zone to the east or both. Because of the major structural boundary just to the east of the 1946 aftershock zone, and the more recent occurrence of the 1946 and 1957 events we regard rupture into the Unalaska Gap as less likely than rupture into the 1938 zone which ruptured earlier, more than 40 years ago.

Eastern Gulf of Alaska Tectonics, Seismicity and Volcanism.

The following pages represent excerpts from a recently completed study on the tectonics and seismicity of the Eastern Gulf of Alaska. We restrict ourselves here to quoting Abstract, Introduction, and Conclusion of this study with one exception: in addition we present the section on volcanic features because of their possible future potential for any geothermal energy development in this important region.

TECTONIC MODEL AND SEISMIC POTENTIAL OF THE EASTERN GULF OF ALASKA AND YAKATAGA SEISMIC GAP^{1,3}

By

Omar J. Perez² and Klaus H. Jacob

Lamont-Doherty Geological Observatory of Columbia University

Palisades, New York 10964

ABSTRACT

A new model of the Pacific - North American plate boundary in the eastern Gulf of Alaska is presented. This model is based upon 13 new fault plane solutions, as well as previously published seismological, geophysical and geological data. The model provides a simple interpretation of the superficially complex transition zone between the transform fault

Figure 1. Aftershock areas of earthquakes of magnitude > 7.4 in the Aleutians, southern Alaska and offshore British Columbia from 1938 to 1979, after Sykes [1971] and McCann et al. [1979b]. Heavy arrows denote motion of Pacific plate with respect to North American plate as calculated by Chase [1978] and Jordan and Minster [1978]. Two thousand fathom contour is shown for Aleutian trench. M_s and M_w denote magnitude scales described by Kanamori [1977]. Shumagin gap is in center of figure.

Figure 3. Location/tectonic map of southern Bering Sea, central and eastern Aleutian Islands, Alaska Peninsula, southwestern and southcentral Alaska and western Gulf of Alaska. Principal tectonic features are: (1) Central and eastern Aleutian Arc with trends of Aleutian Islands, Alaska Peninsula, Cook Inlet, active volcanoes (asterisks), and Aleutian Trench (close hatchuring); (2) Bering Shelf; its western margin is outlined by 400 and 2800 m isobaths (all bathymetric contours in meters); (3) Shelf-edge basins, i.e. Navarin and St. George Basins on the Bering Shelf and six forearc basins along the Shumagin-Kodiak Shelf, the first four, from west-to-east, are the Sanak, Shumagin, Tugidak and Albatros Basins. Shelf-edge anticlines associated with some basin are shown by trend lines crossed by opposing arrows; (4) Seamounts of the western Gulf of Alaska (closed isobaths); (5) Major faults (heavy, solid lines, dotted where inferred); some are labeled with letter codes, the first letter corresponding to first letter of fault name, the last being "F" for fault. Specific faults referred to in text are the Bruin Bay Fault (BBF) and the informally named Kodiak-Montague Island Fault system (K-MIF). Place names used in text and then codes are: Becharof Lake (BL) and Wide Bay (WB).

SHUMAGIN ISLANDS EPICENTERS

Figure 4. Epicentral map from local network data. Epicenters in the Shumagin area are shown for the period July, 1973 through June, 1977 and April, 1978 through December, 1978. All epicentral locations satisfy the following criteria: (1) rms arrival-time residual less than 0.21 sec; (2) vertical and horizontal error estimates less than 36.3 and 28.3 km, respectively; (3) number of stations greater than or equal to 3; and (4) number of phase arrivals greater than or equal to 5. Depths are coded as follows: open circle, 0-20 km; open square, 20-60 km; "x", 60-210 km. Stations of the Shumagin Network are shown by solid triangles. Water depth is shown by isobaths in fathoms. Surface locations of hypocentral cross-sections are marked by dashed rectangles. Along the strike of the arc, the width of these sections is about 150 km; "W", "C", and "E" mean western, central, and eastern, respectively. The central section is divided into two additional cross-sections, "CW" and "CE", (central-west, and central-east) that are about 75 km wide. These sections are shown in Figures 5 and 6.

SHUMAGIN ISLANDS CROSS SECTION

17e

Figure 5. Topographic and hypocentral cross-sections through the Shumagin Islands. All distances and depths in these figures are given in km. The topographic section (top) has a vertical exaggeration of 10:1. The hypocentral section (bottom) is a view looking ENE through the volume indicated by the rectangle labeled "C" in Figure 4. The topographic section is a generalization of the relief along the axis of this rectangle; that is, from the Bering Sea, across the Alaska Peninsula, seaward through the Shumagin Islands, the outer shelf, and to the Aleutian trench. In both sections NNW is to the left. The approximate positions and depth to basement of the sediment-filled forearc basins are indicated by a dashed line. This line is based on a rough average of depths to basement in cross-sections of the Sanak, Shumagin and Tugidak Basins (Bruns and von Huene, 1977 and Fisher, 1979). Station locations of the Shumagin Seismograph Network are shown by solid triangles along top of the hypocentral section. Designator codes for stations are plotted above triangles. Hypocenters are shown by open squares. For quality criteria of hypocenter data, see caption of Figure 4, and text.

Figure 6. Additional hypocentral cross-sections from the Shumagin Islands area. The four sections shown are plotted in the same manner as the section shown in the bottom part of Figure 5. The locations of these sections are given by the dashed rectangles in Figure 4. The sections "CW" and "CE" are the western and eastern halves, respectively, of section "C" shown in Figure 5. Sections "W" and "E" are adjacent to "C" on the west and east, respectively. For discussion of apparent larger scatter in hypocenters in the two bottom sections (i.e. "W" and "E") see text.

Figure 7. Epicentral plot of teleseismically located earthquakes shallower than 80 km in the eastern Aleutians and southern Alaska that occurred between 1965 and June 1979. Only earthquakes assigned magnitude 5.0 or greater are included. Epicenters are from the Preliminary Determination of Epicenters (PDE) of the U.S.C.S. Symbol size is scaled to magnitude shallow earthquakes (depth ≤ 60 km) are indicated by a circle, earthquakes at depths between 60 and 80 km are indicated by an x. Bathymetric contours are in meters.

Figure 8. Epicentral plot similar to Figure 7, except of earthquakes that occurred between 1945 and 1964. Epicenters are from the relocations of Tobin and Sykes [1966], Sykes [1977] and from the International Seismological Center and the U.S. Coast and Geodetic Survey (PDE). Symbols are as Figure 7.

Figure 9. Compilation of mostly new or unpublished fault-plane solutions for the eastern Aleutian arc. For parameters and references regarding these solutions see Table 1. Lower-hemisphere equal-area projections of the focal spheres are shown. Solid black quadrants represent compressional P arrivals at stations, open quadrants dilatational arrivals. Small open circle in the compressional quadrant represents T axis, small solid circle in the dilatational quadrant represents P axis. Note the following type of solutions: #1 and 2 shallow normal faulting; #3 and 4, shallow strike-slip faulting; #5 - 13, shallow thrust faulting; #14 - 18, intermediate-depth down-dip tensional solutions; #19 - 21, intermediate-depth down-dip compressional solutions. Also note transition in the overriding North American plate from oceanic Aleutian Basin west of 400 m isobath to continental Alaska-Bering Sea margin east of this isobath. This transition coincides with the change from down-dip compressional solutions in the western portion to down-dip tensional solutions in the eastern portion of the descending slab. Solid triangles are active volcanoes (large symbols) and Quaternary volcanoes (small symbols), respectively.

tectonics of British Columbia and the subduction zone tectonics of the eastern Aleutian arc. The southern end of this zone, dominated by strike-slip motion (at a rate of 5.5 cm/yr) on the Queen Charlotte - Fairweather fault system, is shown to have a 1.3 cm/yr component of convergence. This small component of convergence results in highly oblique subduction of Pacific lithosphere, and arc volcanism. Along the northern end of the transition zone plate convergence dominates, with rates of 6.0 cm/yr near Yakutat Bay to 6.3 cm/yr off Kayak Island. This part of the zone is arcuate; the major faults and thrust sheets between them have an onion-like structure. The zone of deformation caused by the convergence, which includes the Chugach - St. Elias Range, is about 500 km wide. Seaward of the deformed zone is the "Yakutat Wedge", a semi-allochthonous slice of continental shelf. The Yakutat Wedge is being underthrust by Pacific lithosphere, along a northeast dipping, shallow-angle thrust zone. About 1.2 cm/yr of the total convergence is taken up by this thrust zone, while 5.0 cm/yr is taken up by the Chugach - St. Elias Range. Minor convergence (a few mm/yr) may be taken up by distributed crustal shortening as far north as the Alaskan Range. Geologic evidence shows only small Recent motion on the Denali, Totschunda and other inland faults.

The model implies that the Queen Charlotte - Fairweather - Chugach - St. Elias fault systems have the greatest seismic hazards. Recurrence times are probably between 50 and 200 years, with a temporarily very high potential for a great earthquake in the Yakataga seismic gap. Earthquakes beneath the Yakutat Wedge present more moderate seismic hazards, having recurrence times for great events on the order of several hundred years or more. Inland faults are probably capable to produce large earthquakes; such events are estimated to be infrequent; their recurrence intervals

cannot be accurately determined but may be many hundreds or thousands of years.

INTRODUCTION

The Gulf of Alaska (Figure 10) has been the site of large and great historic earthquakes. In the past, seismicity in this region caused relatively little damage to people, because of the region's sparse population and low level of economic development. Today, however, its economic interests and population are rapidly growing. Petroleum exploration off the continental shelf of the Gulf of Alaska has begun. Consequently, the need to understand the structure of this region, and the need to reliably estimate seismic hazards have both increased. The region provides an important testing ground for earthquake prediction techniques.

We present a new model for the present tectonics of the Gulf of Alaska, a region of transition from predominant transform motion in the southeast to subduction tectonics in the northwest, involving the Pacific and North American plates. We are able to bring together a variety of seismological, geological and geophysical data into a coherent, self-consistent framework. Some of these data, in particular thirteen fault plane solutions, are new; others are taken from the literature (Figures 11 and 15, Table 2).

Our model has severe implications for assessing seismic hazards in southern Alaska. For instance, we conclude that the area called the 'Yakataga seismic gap' can be expected to rupture in a single great event or several large earthquakes in the not too distant future. This finding calls for intensified studies towards earthquake prediction in this gap.

A major tectonic finding is that the transitional tectonics of the Gulf of Alaska extends farther southward than heretofore recognized. Some limited convergence, in the form of lithospheric subduction (Figure 12), takes place obliquely (almost parallel) to the Fairweather - Queen Charlotte faults, usually taken as a classical transform-fault system. This situation may have wide-ranging implications for slowly convergent transform faults in other parts of the world and may be the key for understanding how subduction is initiated along a certain class of ocean-continent plate boundaries.

Volcanic features - and the descending Pacific lithosphere. Landward of the seismically defined Pacific-North American plate boundary one finds in Southcentral and Southeast Alaska and in the adjacent Canadian territories two types of Late Tertiary to Quaternary volcanic features: (1) a mostly andesitic, calcaikaline volcanic arc in a predominantly compressional tectonic regime; (2) scattered appearances of alkaline to peralkaline olivine basalts in either extensional back-arc regions or in regions of leaky transform faulting.

The significance of these volcanic features lies in their potential to shed light on the tectonics of regions where seismologic and other geophysical and geologic evidence is either absent, sparse, or inconclusive.

The most obvious volcanic feature (Figure 12) is the broad arc of either active, Quaternary or Late Tertiary volcanoes in the Wrangell Mountains which continues less conspicuously to the southeast into Canadian territory where Souther (1977) reports their age mostly Late Tertiary, except for two possibly younger, small basaltic centers in the headwaters of the Alsek River (schematically indicated near 60°N, 137°W;

Figure 12). The Wrangell lavas are calcalkaline and hence strongly suggest a relationship to a subducted oceanic lithosphere beneath them (Figure 13). Whether Mt. Edgecumbe volcano (57°N , 136°W ; Figure 12) near Sitka is part of the same volcanic arc is highly questionable since the volcanic front associated with calcalkaline volcanoes is commonly found above or slightly landward of the 100 km depth contour of the associated dipping seismic zone, where any has developed. If a subducted oceanic lithosphere would exist beneath Mt. Edgecumbe volcano, it would have to dip very steeply with a sharp bend from near-horizontal, offshore of the continental shelf in the west, to near-vertical beneath and east of the Fairweather fault. Mt. Edgecumbe lies only about 30 km to the east of the Fairweather fault, which is offshore at this latitude.

The speculation that a Pacific lithosphere may dip beneath most of the northern half of the Queen Charlotte-Fairweather fault system (Figure 14), is strengthened by a kinematic consideration: recent plate-tectonic models such as model RM2 (Minster and Jordan, 1978) show a small but finite component of convergence across the Queen Charlotte-Fairweather fault system which averages to 1.3 cm/yr for the fault system between Yakutat Bay in the north and the Pacific, North American and Juan de Fuca plates triple junction near Vancouver Island in the south.

The depth to which the slab of Pacific lithosphere may extend probably increases from south to north as the age of Pacific ocean floor increases along the margin (Atwater, 1970). Mt. Edgecumbe volcano is located at a northern latitude more than half the distance between the triple-junction and Yakutat Bay, and the age of the leading edge of Pacific lithosphere beneath Mt. Edgecumbe volcano can be estimated from magnetic data (Pitman and Hayes, 1968; Atwater and Menard, 1970; Naugler and Wageman, 1973; Taylor and O'Neill, 1974) to be approximately 10 MY old ($\sim$ anomaly 5).

If the present rate of convergence (1.3 cm/yr) is extrapolated back in time one would require a period on the order of 10 MY to subduct the Pacific lithosphere beneath Mt. Edgumbe volcano to a depth (of about 100 km) where andesitic volcanism can commence. The time required could be even longer if Atwater and Molnar (1973) are correct that prior to about 4.5 MY the Pacific-North American plate motion was slower. At latitudes south of Mt. Edgumbe volcano andesitic-type volcanism is less likely to occur since the ocean floor becomes younger towards the triple junction and may not have been available for a sufficiently long period to be subducted to depths required for producing andesitic-type volcanism.

In fact, south of latitude 58°N, throughout the entire Alexander Archipelago from Chatham Strait to Vancouver Island, and in adjacent coastal British Columbia, one finds another mode of Late Tertiary to Quaternary volcanism prevailing. It appears to be associated with extensional tectonics. The Stikine volcanic belt (Souther, 1977) is composed of more than 50 post-glacial eruptive centers and at least as many of Late Miocene to Late Pliocene age. This belt trends northerly, sub-parallel to Chatham Strait fault, more than 200 km east of the latter. The lavas are mainly olivine basalt, and the Quaternary centers have been formed by monogenetic (single-pulse) activity. The monogenetic activity and the chemical composition are exactly the type of volcanism which Nakamura et al. (1977) have elsewhere in Alaska identified to represent extensional tectonics. Souther (1977, his Figures 5 and 6) gives an identical explanation for the Stikine Belt lavas and associated volcanic features. He points out that they are aligned such that their east-west directed component of rifting is compatible with the tensional axis associated with the stress system of the right-lateral Queen Charlotte transform fault.

In general all active right-lateral faults with an orientation slightly rotated clockwise from the north-northwest oriented slip vector of regional plate motion (i.e. to the north or northeast) should show a component of opening, while those with counterclockwise rotated orientation (i.e. northwest) should show a component of convergence as described earlier for the Fairweather and Queen Charlotte faults. Apparently, the two modes can coexist to a certain extent such that seaward and near the Queen Charlotte fault - i.e. on the frontal side of this slow 'subduction zone' a small component of convergence is present, while to the east of it in the 'back-arc' region a component of opening is present in some locations.

It appears that south of Cross Sound, the margin of the North American plate (east of the major transform faults) experiences a certain amount of rifting, while north of Cross Sound the compressional tectonics starts to dominate. This is not only manifested in the type of volcanism but also in the topography indicating tensional block-tectonics with local horst and graben formation in the Alexander Archipelago, while to the north crustal shortening and strong uplift commence at the southern tip of the Fairweather Range and increase northwestward towards the Chugach-St. Elias Range.

We recognize the limitations of our arguments for a slowly subducting slab beneath the Queen Charlotte-Fairweather fault system which has been heretofore thought of mostly as a classical transform region. These limitations stem largely from the uncertainty in the pole position for the present relative motion of the Pacific-North American plates (Minster and Jordan, 1978); even larger uncertainties may be involved when such motions are traced back to about Late Oligocene times (Atwater and Molnar, 1973).

The role of the Farallon and Kula plates must also be considered for those epochs. Another complication may arise from our poor understanding of intraplate deformation of western North America and of portions within the leading young edge of the Pacific plate. All these problems were alluded to by Atwater (1970). A model allowing for minor subduction of Pacific ocean floor beneath the Fairweather and Queen Charlotte faults seems, however, to emerge partly on grounds of volcanologic data and appears worth testing in more detail.

CONCLUSIONS

Despite the apparent complexity of the tectonics of the eastern Gulf of Alaska, we have attempted to construct a relatively simple tectonic model. This model describes the gross geologic features of about the last 10 MY, but extrapolation to 20 MY may be possible without major changes. This model is consistent with globally computed rates and directions of motion between the Pacific and North American plates, but contradicts some - yet not all - of the recent inferences which some investigators of marine - and land - geological data have made.

The model indicates north-northwesterly motion of the Pacific plate along the Queen Charlotte-Fairweather fault system at rates increasing from about 5.5 cm/year in the south to 5.9 cm/year in the north. The sense of motion is predominantly right-lateral strike-slip but contains a component of convergence which when averaged across the entire length of fault-system, amounts to about 1.3 cm/year. The pure strike-slip rates vary from 5.3 cm/year in the south to 5.7 cm/year in the north. The small convergence - when active over several million years - causes a significant

amount of subduction of Pacific oceanic lithosphere beneath the western continental margin of British Columbia and Southeast Alaska; transform faulting predominates, however, and results in probably more than 500 km of lateral displacements during the last 10 MY.

The depth to which the leading edge of Pacific lithosphere has been thrust beneath this active continental margin increases from south to north; for latitudes north of Mt. Edgacumbe volcano, near Sitka, Pacific lithosphere seems to have reached depths in excess of 100 km, sufficient to give rise to calcalkaline magmas which form a poorly developed, discontinuous volcanic arc above the dipping slab. To the northwest, the arc extends into the volcanic Wrangell Mountains. Northwest of the Wrangell volcanoes the arc abruptly terminates near the Copper River valley for yet unexplained reasons. No intermediate-depth seismicity is associated with the obliquely subducting portion of Pacific lithosphere, apparently because of the lithosphere's young age and slow rate of subduction.

In the northernmost part of the Gulf, west of Yakutat Bay, the plate boundary bends westward approximately following the coast line. There the component of convergence becomes predominant over the transform component. The total relative plate motion is of the order of 6 cm/year increasing to 6.3 cm/year near Kayak Island. A small portion, probably about 20%, of this motion is taken up by thrusting along the off-shore shelf-edge structure that seaward delimits the Yakutat wedge; by this motion the sedimentary wedge overrides the Pacific ocean floor in a southwestern direction.

The much larger portion (probably close to 80%) of Pacific-North American plate motion is absorbed in a roughly 100 km wide thrust belt (measured across strike), the most prominent expression of which is the Chugach-St. Elias range. This active thrust belt extends seaward probably

as far as the Pamplona zone, and inland to about the Border Ranges fault. The most intensive deformation, i.e. crustal shortening and associated uplift, occurs directly above the rupture zones of most large and all great earthquakes that have occurred here during historic times.

A major seismic gap has been identified to extend along strike of this thrust belt from about Icy Bay (west of Yakutat Bay) to about Kayak Island. The region of this seismic gap is estimated to have stored sufficient strain energy since it last ruptured in 1899 to release an average slip of several meters over an area of at least 150 by 100 km. This strain would be sufficient for at least one great earthquake. The timing for such a great event, or - instead - a sequence of large events, is uncertain and could be either imminent or a few decades away. This region is certainly a prime target for testing earthquake prediction efforts.

A major result of this study is that the transition from the transform plate boundary along British Columbia to the convergent plate boundary at the eastern Aleutian trench is much more gradual than heretofore recognized. The geological importance of a small component of convergence at a predominant transform fault along an ocean-continent boundary needs further evaluation. If a substantial component of convergence persists for several million or tens of million of years it apparently can lead to, first, initiation of subduction, and later to a sufficient amount of subduction of ocean floor that causes tectonic features not unsimilar to those of active convergent margins. The importance of such features may become more evident when they are compared to those of other boundaries with histories of combined strike-slip and thrusting motion, i.e. in southern Chile, New Zealand, the Sumatra-Andaman arc, or even parts of coastal California straddling the San Andreas fault. Such comparisons may provide

TABLE 2. Eastern Gulf of Alaska Focal-Mechanism and Earthquake Parameters

No.	Date	Lat. ^d (°N)	Long. ^d (°W)	h ^d (km)	mb ^d	(M _S) ^d [M _W]	Plane 1**		Plane 2		P Axis		T Axis		B Axis	
							Strike Azi.	Dip	Strike Azi.	Dip	Trend (°)	Plunge (°)	Trend (°)	Plunge (°)	Trend (°)	Plunge (°)
1	8/15/72	56.31	135.57	23	5.4	(4.8)	344	90	74	90	209	0	299	0	-	90
2	7/30/72	56.77	135.91	29	6.2	(7.6)	344	73°N	72	83°S	207	17	299	7	50	73
3*	24/10/27	57.69	136.07	0		(7.1)	329	68°N	64	78°S	195	7	289	24	90	65 ¹
4*	10/07/58	58.34	136.52	0		[8.2]	335	72°N	68	82°S	200	6	293	20	96	70 ²
5	7/01/73	57.86	137.42	15 ⁺	6.2	(6.7)	295	24°N	301	66°S	210	21	35	65	300	2
6	7/03/73	57.99	138.04	19 ⁺	6.1	(6.0)	295	24°N	302	66°S	210	21	37	69	301	2
7	7/05/73	57.90	137.99	32	5.4	(4.9)	298	24°N	300	66°S	210	21	33	69	300	1
8 [§]	2/28/79	60.62	141.50	11		(7.7)	70	10°N	75	80°S	164	35	344	55	255	1
9	12/23/65	60.57	140.64	25	5.7	(5.8)	314	15°N	85	80°S	185	34	342	54	90	11
10	3/26/71	60.33	140.94	15 ⁺	5.8	(5.8)	277	10°N	82	80°S	178	34	338	54	82	10
11	4/16/70	59.84	142.43	7	5.6	(7.0)	295	12°N	95	78°S	188	33	358	57	96	4
12	4/19/70	59.60	142.72	20	5.6	(6.0)	322	9°N	79	85°S	182	35	357	55	80	12
13 [†]	5/17/64	59.40	142.70	35	5.1		329	15°N	102	80°S	201	34	360	54	104	10 ³
14 [†]	3/30/64	59.90	145.70	15	5.6		275	6°N	310	84°S	215	40	43	50	309	3 ³
15	5/18/69	60.34	145.93	6	5.4	(5.2)	319	12°N	90	82°S	188	36	349	52	91	9
16	8/18/70	60.70	145.38	30	5.8	(5.9)	348	15°N	92	86°S	197	38	347	47	93	15
17	9/06/73	61.04	146.85	17	5.5	(5.5)	72	18°N	60	72°S	153	27	325	63	62	5
18 ^{††}	3/28/64	61.10	147.60	20	8.5	[9.2]	---	---	66	85°S	---	---	---	---	---	---
19 [†]	3/28/64	59.80	148.70	40	5.8		24	7°N	72	85°S	158	40	348	50	252	5 ³

^d ISC data, except where other source is indicated.[M_W] After Kanamori [1977].

† Aftershocks of the 1964 great Alaska event. Locations are from NOAA file.

†† Great Alaska event. Relocated by Sykes [1971].

* Relocated by Tobin and Sykes [1968].

+ Depth obtained from pP readings on WWSSN records.

§ St. Elias event. Location and magnitude after Lahr et al. [1979]. Depth after Boatwright and Perez [1979].

1 Stauder [1959]; 2 Stauder [1960]; 3 Stauder and Bollinger [1966].

** Selected plane of faulting.

Figure 10. Map showing location of major faults with Late Cenozoic displacement (solid lines) in southeast Alaska and adjacent portions of Canada (King, 1967; Plafker *et al.*, 1978). Dashed lines: faults with possible late Cenozoic motion. Shaded areas denote off-shore zones of active thrust faults and folds. Broken shading indicate possible faults suggested by patterns of shallow seismicity. Long arrows represent motion of Pacific plate with respect to North American plate at different localities (Chase, 1978; Minster and Jordan, 1978); rates are given below arrows. Stars denote volcanoes of Quaternary age. MIs, Montague Island; PWS, Prince William Sound; CRD, Copper River delta; KIs, Kayak Island; IB, Icy Bay; YB, Yakutat Bay. The Gulf of Alaska Tertiary province is interpreted as formed by a series of sedimentary wedges subducted by Pacific Ocean floor. The subduction commences at the northwest trending off-shore fault zone (shelf-edge structure) which connects the northern tip of the Queen Charlotte fault to the eastern Aleutian trench south of Prince William Sound.

Figure 11. Focal mechanisms and major faults in southeast Alaska and in adjacent parts of Canada. The number for each solution is indicated near its epicenter (solid dots). Shaded areas in the focal mechanisms (lower hemisphere projections) represent compressional first motion; P and T denote inferred axes of maximum and minimum principal stress. Triangle and stippled area indicate epicenter and rupture zone (Lahr *et al.*, 1979a) of the $M_s = 7.7$ St. Elias earthquake of February 28, 1979. Pamplona fault zone, PZ, indicated by area between small arrows. Yakutat wedge, YW, is bounded by Pamplona zone, Fairweather fault (FF) and the off-shore fault zone of the shelf-edge structure approximately between events 5 and 13. PWS, Prince William Sound; KIs, Kayak Island; YAK, Yakataga district; IB, Icy Bay; YB, Yakutat Bay; CS, Cross Sound; S, Sitka. Stars denote Wrangell volcanoes, WV, of Quaternary age. Faults: D-F, Denali; T-F, Totschunda; CH-STE-F, Chugach-St. Elias; CH-ST-F, Chatham Strait; QCH-F, Queen Charlotte. Heavy arrow represents motion of Pacific plate with respect to North American plate (Chase, 1978; Minster and Jordan, 1973). Submarine contours in meters. For solutions 1 through 4 the northwesterly striking nodal plane is chosen as rupture plane, for all other solutions the shallow dipping plane is assumed to represent the rupture zone.

Figure 12. Slip-vectors obtained from fault-plane solutions of shallow earthquakes in southern and southeast Alaska (solid arrows). Solutions 1 to 19 are from focal mechanisms in Figure 11; solutions 20 to 23 are from Stauder and Bollinger (1966). Hollow arrows represent calculated directions of Pacific-North America plate motion at the same localities as earthquakes 1 to 23 (after global fit of Minster and Jordan, 1978). Stars represent andesitic volcanoes of Quaternary age. Queries indicate uncertain occurrences. Note the agreement between direction of slip vectors derived from fault-plane solutions and the calculated directions of plate movement. Exceptions are mechanisms 5 to 7 and 13 and 14. See 'Tectonic Model' section in text for their interpretation.

Figure 13. Highly schematic cross section (no vertical exaggeration) across the Chugach-St. Elias Mountains in the Gulf of Alaska. This section trends due north from a point offshore at 58.82°N, 142.2°W (0 km at right-hand corner) to a point inland at 63.32°N, 142.20°W (500 km at left-hand corner). Geologic details and crustal thicknesses are mostly inferred. This cross section illustrates schematically the inferred model of the imbricate thrust system and internal deformation of the wedge-shaped leading edge of the North American plate as it overrides the downgoing Pacific plate. The ratio of convergent to transform motion in this section is approximately 5 cm/year to 1 cm/year (see arrow symbols). The amount of convergence between any overriding thrust slice and the underriding Pacific plate is inferred to increase landward where beneath the Chugach-St. Elias Range it may reach a maximum. Minor crustal deformation in the overriding plate may extend as far north as the Alaska Range. No intermediate-depth earthquakes exist to delineate the downgoing slab of Pacific lithosphere. Its position is inferred from the Wrangell volcanoes. The North American plate consists of several distinct allochthonous terranes (e.g. Wrangellia) which since the Mesozoic moved into place along major strike-slip faults. Bouguer gravity anomalies Δg_B (in milligals), after Barnes (1977).

Figure 14. Highly schematic cross section (no vertical exaggeration) across the Shelf Edge Structure and the Fairweather fault and mountain range. This section trends N50°E starting from a point offshore at 57.9°N, 139.2°W (10 km at right-hand corner) to a point inland at 59.5°N, 135.0°W (300 km at left-hand corner). The ratio of convergent to transform motion in this section is approximately 1 cm/year to 5 cm/year (see arrow symbols). The fault with the presently highest rate of slip is the steeply dipping Fairweather fault taking up at least 5 cm/year of right-lateral slip as indicated. We infer that this shear zone merges at a depth below 20 to 40 km with the subduction zone. Slip vectors with numbers 1 to 4 and 5 to 7 represent fault plane solutions (compare Table 2 and Figures 11 and 12). The depth to which Pacific lithosphere has descended is uncertain but is probably in excess of 100 km. The existence of a former Farallon plate at the indicated location in the mantle is highly speculative and may apply only for geological periods some 10 to 20 MY ago. The volcanic arc in this section is very poorly developed. For allochthonous terranes (i.e., Wrangellia), and Bouguer gravity data see legend of Figure 13.

Figure 15. Compilation of the thirteen fault-plane solutions newly derived for this paper. The solutions are numbered in accordance with those of Table 2 and as used throughout this paper. Lower-hemisphere equal-area projections are shown. Solid dots are for compressional, open dots for dilatational, and X's for nodal P arrivals. Most polarities are from reliable readings of long-period records (large symbols); a few short-period readings (small symbols) are used, but are considered unreliable. Short lines indicate S-wave polarizations. Solid nodal planes are well constrained, broken ones poorly constrained. P = inferred axis of maximum principal stress, T = axis of minimum principal stress. Except for solutions 1 and 2 which show mostly strike slip, all other solutions show mostly dip slip with faulting inferred to occur on the shallow dipping plane.

the key to understanding how subduction of very young ocean floor resumes after passage of a triple junction; i.e. after consumption of an oceanic spreading center into a trench system. It appears that at these times of plate-boundary readjustments transform faulting combined with some convergent plate motion is typical; large terranes can be transformed large distances along the continental margin and may sometimes undergo substantial rotations during this process. The Yakutat wedge could be the latest and youngest such example of translation of an allochthonous terrane, in this case along the Fairweather fault. By its crustal structure the wedge may be only a rootless sedimentary thrust slice, rather than a fully developed continental block, and hence should not be considered a mature miniplate.

Pavlof Volcano Seismicity

Lamont-Doherty operates 14 seismometers in the immediate vicinity of Pavlof Volcano. This is one of the densest seismic networks in existence at a volcano. Twelve instruments are part of the local Pavlof network installed by Lamont scientists in 1976. These are labelled PN1-8 and PS1-4 on Figure 16. Data from these DOE-funded instruments are being used to delineate a magma chamber beneath Pavlof Volcano using seismic wave attenuation and S-wave screening techniques. Magnetic tapes have been played back to produce paper records for about one half of the data, and we are currently trying to erase our backlog. Preliminary analysis has been carried out on a small number of events. We are re-thinking our approach to the problem in light of some promising results from other work in progress.

Two other stations, PVV and BLH (Black Hills; also Figure 16) are part of the larger Shumagin Islands Network. PVV is operated in conjunction with the University of Alaska. Paper helicorder seismograms have been recorded on a fairly continuous basis from the PVV station since 1973. These paper records are an important data source because they allow us to overview at a glance the seismic activity of the volcano. All of the helicorder records are now completely scanned for determining the quality and extent of the data. Pavlof has erupted in fall 1973, fall 1974, fall 1975, fall 1976 and fall 1977, and not in 1978 or 1979. Helicorder records have been analyzed for the 1973 and 1974 eruptions: daily event counts are summarized in histograms (Figures 22 and 23).

A detailed study has been made of the Fall 1974 eruption sequence for which the data set is complete and of high quality. The number of volcanic events of each of four different types were counted for each 2-hour interval for the entire period from September 1, 1974 to April 30, 1975. The event types are: high-frequency tremors (Figure 17), harmonic tremor (Figure 18), explosion earthquakes (Figure 19), and B-type earthquakes (Figure 20). An example of a histogram representing the detailed study is shown in Figure 21 (note that no high-frequency tremors occurred during this time period). Part D of Figure 21 shows a theoretical earth-tidal strain time series calculated for Pavlof. The azimuth of 135° is approximately the azimuth of convergence of the North American and Pacific plates. Note the correlation between B-type earthquake swarms and the earth tide.

In the Appendix we include an abstract of a paper presented at a special conference on volcanism during the Fall 1979 Pacific Northwest American Geophysical Union Meeting. The relationship between the volcanic

earthquake swarms and the earth tide is described more fully in this abstract. A paper describing this work in detail is currently being prepared for publication.

This work constitutes a minor breakthrough in volcanology. Many previous papers, for example, Hamilton (1973), Johnston and Mauk (1972), and Ryall et al. (1968) have dealt with earth tidal triggering of both volcanic eruptions and tectonic earthquakes. Only one widely cited paper (Mauk and Kienle, 1973) has dealt with Earth tidal triggering of volcanic earthquakes. This paper has been shown to be wrong (Doug Lalla, 1979, personal communication). Thus the work in progress at Pavlof may be the first to show the relationship between the solid earth tide and volcanic earthquakes. From our studies, we conclude that Pavlof volcano is much more sensitive and responsive to small changes in stress than was previously thought. The earth tides, with amplitudes in the range from 10^{-2} to 10^{-1} bar, have a direct effect on seismicity.

From this information we have been able to measure in rough terms the rates of near-surface activity, i.e. it takes magma 3-4 days to make its final ascent to the surface before an eruption, and conversely 3-4 days to make its descent following the eruption. We are presently engaged in a study of the spectral characteristics of harmonic tremor in an effort to learn more about the dimensions of the magma conduits. The studies converge in that we presume the magma makes its final ascent to the surface from a magma chamber. By determining in a rough way the depth to such a body, we will be in a better position to resume data reduction and analysis of the Pavlof network data. Obviously we will concentrate our efforts on the data pertaining most closely to the zone or zones inferred from the studies mentioned above.

We expect that attenuation and screening of teleseismic waves would change corresponding to periods of magma movement. Thus it is very important to establish beforehand just what stage of activity the volcano is in so that locating a magma chamber or zone of attenuation will be meaningful. The zone of attenuation would be largest during periods of magma movement and eruption; this is the time when the volcano is in an inflated state (Fiske and Kinoshita, 1969). When the volcano is inactive we would expect to locate the smaller, more permanent part of the magma chamber.

The preceding paragraphs explain why we feel it is necessary to analyze the PVV records first before undertaking the bulk of the data reduction and analysis for our magma chamber search. We feel that much important information concerning patterns of eruptive behavior will be gained from a thorough and systematic study of the PVV records. We are about one third of the way through analyzing these records.

We are also, however, presently making use of the Pavlof network data in several important auxiliary ways. One way is that data gaps appear in the PVV coverage. We use the Pavlof network stations to fill in the existing gaps, where possible. A second way is that as stations in the Shumagin array fail, due to Aleutian weather, bears or other field or technical conditions, we substitute the Pavlof network stations onto the develocorder. This does not affect the normal recording of Pavlof data on magnetic tape, but it greatly enhances the quality of our earthquake locations at season's end when we are often down to our last few stations. Some of the Pavlof network stations were, in fact, recorded on the develocorder during the first half of 1979.

Figure 16. Location map showing seismometer stations around Pavlof Volcano. PVV and BLH are part of the Shumagin Islands Regional Network. PN1-8 and PS1-4 are stations of the Pavlof Local Network.

Figure 17. High frequency tremors as they appear on the PVV heliograph records. The tick marks are one minute apart. This swarm of high frequency tremors took place on October 28, 1974 just a few hours before Pavlof volcano began to erupt.

Figure 18. Harmonic tremor as it appears on the PVV helicorder records. This signal occurred on November 17, 1974, at 1100 hrs, GMT. The signal duration was approximately 22 minutes. Note the dominant frequency of 1.5 Hz. Spectral analysis is in progress to determine the total frequency content of the signal.

Figure 19. Explosion earthquakes as they appear on the PVV helicorder records. The arrows point toward the air shock phase, which shows up on the seismogram 21 seconds after the ground phase. These explosion earthquakes indicate when the volcano is actively erupting.

Figure 20. B-type earthquakes as they appear on the PVV records. The dominant frequency of these events are 1.5 Hz, which is the same dominant frequency observed for harmonic tremor. B-type earthquakes occur in swarms just before, during and just after eruptions at Pavlof.

PAVLOF VOLCANO SEISMICITY

A) HARMONIC TREMOR

x = OBSERVED

B) EXPLOSION EARTHQUAKES

C) VOLCANIC EARTHQUAKES

B-TYPE

D) PAVLOF STRAIN AZI 135°

Figure 21. 2-hour histograms of: A) harmonic tremor; B) explosion earthquakes; C) B-type volcanic earthquakes. (No high frequency tremors were observed during this time period). Part D) shows a computer-generated theoretical earth-cidal strain time series. Pavlof erupted explosively from October 28 to November 17, 1974 then subsided until November 25, at which time the explosions resumed. The left side of the figure shows daily b-type earthquake swarms that correlate with the maximum semi-diurnal low tide or compression. The right side of the figure shows b-type earthquake swarms that correlate with the maximum semi-diurnal high tide or extension. This same pattern shows up several times during the fall 1974 eruption sequence.

PAVLOF VOLCANO SEISMICITY

Figure 22. Histogram of the number of volcanic earthquakes per day during the fall 1973 eruption. Note the main peak of over 2000 events per day.

Figure 23. Histogram of the number of volcanic earthquakes per day during the fall 1974 eruption. This eruption was of lower intensity than the fall 1973 eruption. Note that the highest peak represents only about 450 events per day.

A third and major way the Pavlof network data has been invaluable is in the study of explosion earthquakes. As shown in last year's annual report, distinct air phases show up on the seismograms for explosive events; these events also appear on the PVV records (Figure 19). We used the arrivals of the Pavlof network to confirm that these events are explosions by constructing a time/distance plot for these events. This information is extremely important because we use the explosive events as indicators of whether or not the volcano is actively erupting. Thus, in a very crucial way, our work on Pavlof already depends on data from the Pavlof seismic network.

In summary, during the past year we have made significant progress in understanding the causes of patterns of eruptive behaviour at Pavlof Volcano. Those patterns may be linked to the existence of a magma chamber or zone of attenuation at some depth within the volcano. We are studying the patterns of activity first so that insights and interpretations made will benefit our search for a magma chamber. The size and position of the magma chamber and patterns of eruption are important pieces of information needed to assess both volcanic hazards and geothermal power potential, and to increase our overall scientific understanding of volcanoes as well.

P-Residual Study

The unusually large positive P residuals that we reported last year have been re-evaluated, and appear to be much smaller, although still positive. The arrival times at Alaskan stations for all events we studied were read again, and a new, improved program for calculating distances, azimuths, and residuals was used. We have expanded our data base to

include first arrivals to stations in the Gulf of Alaska region, operated by the U.S.G.S. Figure 24 shows the distribution of stations whose data we are currently using. First arrivals are read from develocorder film records, with an accuracy of $\pm .05$ sec. Also, we have standardized our data set more carefully, and compute all residuals with respect to the earth model of Herrin et al. (1968). The same trends in P residuals prevail, but the absolute values are substantially reduced.

Nuclear explosions detonated on Novaya Zemlya, for which the azimuths to stations in Alaska are north to north-northwest, yield positive P residuals which range from .5 sec to 3.5 sec. Residuals from nuclear explosions detonated in Soviet Central Asia are mostly positive, by approximating 1.5 to 3.5 sec. Azimuths for these events are north-northwest. Earthquakes in Central America, in the azimuth range east-southeast to southeast, yield residuals which are approximately 1 to 3 seconds positive. By contrast, the nuclear explosions Cannikin and Milrow, which were detonated on Amchitka Island, yield residuals which are negative by as much as 4 seconds. Rays from these explosions to Alaskan stations have spent a large portion of their total paths within the high-velocity slab which dips beneath the Aleutians (Rowlett and Jacob, 1972, among others). Our results agree with those of Rowlett and Jacob (1972) in that we find the absolute residuals for Cannikin and Milrow to become increasingly negative eastward along the arc. This is consistent with the interpretation that rays are refracted laterally along the slab, and rays which spend more of their total paths within the slab yield more negative residuals. Rays from other azimuths, which spend little or no time within the high-velocity slab, appear to be delayed by other lateral inhomogeneities in the crust and upper mantle in the Alaskan region.

Figure 24. Distribution of stations in Alaska for which we read first arrivals and calculate residuals. It is not always possible to calculate a residual for each station for a number of reasons. Sometimes stations fail during winter months and do not record data; at other times some stations may be so noisy that the first arrivals cannot be discerned and hence residuals cannot be calculated. Also, the distribution of stations varies somewhat from the one shown here over the time period from which we have chosen events.

Figure 25. Residuals for a nuclear explosion on Novaya Zemlya, 9/12/73, 73.32°N, 54.97°E, 06:59:54.6, $M_b = 6.8$. Location taken from I.S.C. catalogue.

Figure 26. Residuals for an earthquake in Central America, 7/13/74, 7.76°N, 77.70°W, 01:18:22.9, $m_b = 6.4$. Location taken from I.S.G. catalogue.

Figure 27. Residuals for Cannikin nuclear explosion on Amchitka Island, 11/6/71, 51.472°N, 179.107°E, $m_b = 6.8$.

NOVAYA ZEMLYA EXPLOSION 9/12/73

Figure 28. Examples of first arrivals read from developocorder records for Novaya Zemlya nuclear explosion, residuals for which are shown in Figure 25. Solid arrow indicates 'picked arrival'; dotted arrow indicates 'expected arrival' based on calculated residual.

CENTRAL AMERICAN EVENT 7/13/74

Figure 29. Seismograms for Central American event, residuals for which are shown in Figure 26. Symbols as in Figure 28.

Figures 25-27 illustrate the distribution of residuals (for arrivals read from the clearest records) for sample events from Novaya Zemlya, Central America, and Amchitka. Figures 28 and 29 show examples of seismograms for the Novaya Zemlya and Central American event, respectively.

Of particular interest is the large negative residual of -4 seconds at Sitka for the Cannikin explosion. Jacob and Perez (in prep.) cite evidence for slow (1-2 cm/yr) convergence between the Pacific plate and what they define as the Yakutat Wedge. They hypothesize that the leading edge of the subducting Pacific plate may reach depths on the order of 100 km in the vicinity of Sitka Island. If this is the case, it may explain the pattern of negative residuals in southeast Alaska from the Amchitka explosions.

We presently prepare a series of computer programs to test this hypothesis by combining three-dimensional ray-tracing and inversion techniques, using methods originally developed by Jacob (1972) and Aki et al. (1977).

PRELIMINARY BODY WAVE ANALYSIS OF THE ST. ELIAS, ALASKA EARTHQUAKE OF FEBRUARY 28, 1979*

By John Boatwright

The analysis of large, shallow-focus earthquakes using body-waves is a difficult undertaking. Body-wave arrivals from these events are complicated by two distinct effects: first, the interference between the direct phase and the depth phases, and second, the generally complex nature of rupture in these events. In their analysis of the 1976 Guatemala earth-

quake, Kanamori and Stewart (1978) resolved these complications using a quasi-deconvolutional approach.

In this paper, we perform a direct deconvolution of five body-wave arrivals from the February 28, 1979, St. Elias, Alaska earthquake, recorded at three long-period WWSSN stations and on the Palisades, New York, broad-band Benioff instruments. This direct deconvolution is made possible by the specific focal mechanism of this earthquake; the technique is not generally applicable to all shallow-focus events. The pulse shape information retrieved includes both moments and rupture durations for the sub-events. An average rupture velocity and rupture direction for the whole event can also be estimated. This analysis provides a unique view of the complexity of rupture in this earthquake, in contrast to the descriptions of rupture determined from aftershock locations and surface-wave analysis.

Short samples (150-300 secs) of the seismograms were digitized and corrected to trace deflection against time by interpolating and correcting for the translation pitch of the recording drum. The corrected traces were bandpassed using a triangle smoothing operating and a second-order high-pass Butterworth filter (corner frequency at either 50 secs or 66 secs), run backwards and forwards over the trace. To obtain the bandpassed ground velocity, the band-passed seismograms were instrument-corrected using a recursive deconvolution derived from a bilinear approximation to the coupled galvanometer-seismometer systems.

Figure 30 shows the fault-plane solution for the earthquake as determined by Perez and Jacob (1979), who concluded on the basis of tectonic considerations that the plane which dips shallowly to the north is the fault plane. This nearly horizontal fault plane insures that the pulse

shapes of the direct phase and the depth phases are nearly identical, and thereby permits the purely impulsive deconvolution.

A low-frequency baseline correction was then subtracted from the deconvolved velocity, giving the "approximate whole space" velocity trace. The AWS velocity was then integrated to obtain the AWS displacement, which has a unipolar displacement pulse shape and a flat baseline. The AWS displacement pulse shapes have three distinct pulses which we have identified as sub-events 1, 2, and 3, preceded by a small initial event, labelled 0 where it can be identified.

Figure 31 illustrates the analysis process for the P wave arrivals at Palisades (PAL). The PAL P-wave is a nodal arrival, and the free-surface operator (FSO) has a somewhat longer and more detailed convolutional inverse. The seismogram as digitized, shown at the bottom of Figure 31, clearly details the initial event, which precedes the sharp onset of sub-event 1 by 7 seconds. It cannot be distinguished, however, in the AWS pulse shapes. Sub-event 1 is substantially stronger at PAL than at other stations studies, while sub-event 3 is weaker. Even more striking is the energetic arrival which separates sub-events 2 and 3. If a variation in the fault-plane solution is responsible for these differences, the fault plane must become somewhat more steep as the rupture proceeds, or the slip vector must rotate slightly counter-clockwise to decrease the amplitude of the sP arrival.

Figure 32 summarizes the conclusions of our analysis. For comparison, the heavy dashed line is the estimate of aftershock area obtained by Lahr et al. (1979) from a U.S.G.S. array of 50 short-period seismic stations. The star is the location of the mainshock epicenter (what we have termed the initial event). Sub-events 1, 2, and 3 are

FEBRUARY 28 1979
60.62°N 141.50°W 11km

Figure 30

P-WAVE AT PAL

Figure 31

Figure 32

plotted with dimensions and spatial separations as estimated from our analysis.

The total bodywave moment estimates from our analysis, 11.6×10^{26} dyne-cm, is a factor of 4 smaller than the surface wave moment estimated by Lahr *et al.* (1979). This discrepancy may result from focusing and amplification of the surface wave energy by a strong velocity contrast in the source region.

INVESTIGATION OF TWO HIGH STRESS-DROP EARTHQUAKES

IN THE SHUMAGIN SEISMIC GAP, ALASKA*

Leigh House and John Boatwright

Lamont-Doherty Geological Observatory and

Department of Geological Sciences of Columbia University

Palisades, New York 10964

ABSTRACT

Two moderate size ($m_b = 5.8, 6.0$) earthquakes occurred within a local network of short-period seismograph stations in the Shumagin Islands, Alaska, on April 6, 1974. They were followed by 69 aftershocks recorded over the next two weeks. Both mainshocks triggered a strong-motion accelerograph (SMA) at Sand Point, 50 km NNW of their epicenters.

High quality locations obtained from local network arrivals for the mainshocks and 29 aftershocks yield depths between 35 km and 45 km and define a plane dipping about 30° to the NW. A nearly pure-thrust focal mechanism for the larger ($m_b = 6.0$) earthquake was obtained from long-period data. The fault plane inferred from this mechanism dips 30° in the

direction N16°W. This sequence was located in the upper portion of the dipping seismic zone beneath the eastern Aleutians and was presumably related to underthrusting of the Pacific plate beneath the North American plate.

Estimates of the source parameters of these earthquakes were obtained from analysis of the SMA data and WWSSN short period data. The WWSSN data indicate that the earthquakes had approximately circular rupture areas. Modelling of the SMA records with a quasi-dynamic source model provides the following estimates of the source parameters for the $m_b = 5.8$ and 6.0 earthquakes, respectively: moments, $M_0 = 3.6$ and 6.6×10^{24} dyne-cm and static stress drops, $\Delta\sigma = 890$ and 650 bars. A high frequency spectral falloff of ω^{-3} suggests that the ruptures stopped gradually.

The Shumagin Islands region is believed to have a high potential for a future large earthquake. The location of this earthquake sequence at what is inferred to be the deepest part of the zone of seismic interplate deformation and the high stress-drops of the mainshocks may reflect a considerable accumulation of stress prior to a major earthquake in the Shumagin Islands region.

CONTINUATION OF SOURCE PARAMETER STUDY

By J. Mori

Using the set of earthquake focal mechanisms compiled by House and Jacob (Figure 9, Table 1), estimates of source parameters for these events are being calculated. These earthquakes form a wide belt stretching from the Eastern Aleutians through the Alaska Peninsula, so they represent stress release in a wide variety of tectonic settings.

The P-arrival of events (with magnitudes $m_b = 5.0$ to 6.5) have pulse shapes with dominant periods on the order of two seconds at teleseismic distances. Such frequencies are well recorded on WWSSN short-period instruments. Therefore, these records provide suitable data for an inversion to determine the size of the fault (a), effective stress (τ_e), stress drop ($\Delta\sigma$), and moment (M_0) of the source.

The method of analysis used is that developed by Boatwright (1979) for investigating a high stress-drop earthquake in the same region (#9, Figure 9). The major steps of this analysis are (Figure 34):

- 1) WWSSN records are researched for well recorded P-arrivals. Stations used are generally 30° - 90° from the epicenter. The few stations which exist closer than 30° generally contain reverberating signals high-frequencies that make the interpretation of these crusts and uppermantle P-waves difficult to use for source-parameter analysis.
- 2) A photographic copy of the record is hand-digitized at a rate of about 10 samples a second. The digitization is processed by a set of computer programs which fits the data to a baseline and band-pass filters it (1.0 to .3 Hz). The low-pass part of the filter is a triangle smoothing operator and the high-pass is a third-order Butterworth filter which is run forward and backward over the data to ensure zero phase shift. Next a bilinear approximation for the instrument response of the WWSSN short-period system is deconvolved from the signal and the resultant signal is a time function of ground displacement. The displacement trace is then differentiated to obtain a velocity plot.

- 3) Assuming the self-similar circular rupture model Boatwright (1979) derives equations for obtaining source parameters from these plots. The moment is given by

$$M_0 = 4\pi(\rho(\xi_0)\rho(x)c(x))^{1/2}c(\xi_0)^{5/2} \frac{R(x,\xi_0)}{F^C(\theta,\phi)} \bar{u}_c(x,0).$$

where $\rho(\xi_0)$ and $\rho(x)$ are densities at the source and receiver, respectively; $c(\xi_0)$ and $c(x)$ are P-wave velocities at the source and receiver, respectively. $R(x,\xi_0)$ is the geometrical spreading and $F^C(\theta,\phi)$ is the radiation pattern. $\bar{u}_c(x,0)$ is the pulse area of the displacement seismogram.

Effective stress, that is the stress drop during rupture, is given by

$$\tau_e = \frac{(\rho(\xi_0)\rho(x)c(x))^{1/2}c(\xi_0)^{5/2}}{v^3} \left(\frac{R(\xi_0,x)}{F^C(\theta,\phi)} \right)^{(1-\zeta^2)^2} \left\langle \frac{\dot{u}_c(x,t)}{t} \right\rangle$$

where v is rupture velocity and ζ is the apparent Mach number defined as $\zeta = \frac{v}{c(\xi_0)} \sin \theta$, where θ is take-off angle measured from the fault normal.

$$\left\langle \frac{\dot{u}_c(x,t)}{t} \right\rangle$$

is the slope of the rupture phase taken directly off the velocity seismogram.

The fault radius is given by $a = T_{1/2} v (1-\zeta)$ where $T_{1/2}$ is the duration of the rupture phase, or equivalently the rise time of

the displacement pulse. The (static) stress drop is given by Brune (1970)

$$\Delta\sigma = \frac{7}{16} \frac{M_0}{a^3}$$

The non-causal part of the processed signal which can be seen as a precursor on the velocity trace, and which becomes greatly amplified on the displacement trace, is a result of the high-pass filter which has a cut-off frequency close to the corner frequency of the P-wave. Unfortunately it is necessary to severely filter the signal in this manner in order to separate a clear unipolar displacement pulse from the slightly longer-period noise prevalent in the short period system.

This analysis has been carried out for four events and the results are summarized in Table 3. There is considerable scatter in source parameters derived for the same event but different stations. Some of this scatter could be caused by the crustal structure near the stations which can greatly change the amplitudes of a signal in the frequency band studied. The fact that many stations are used is a check on the effects of propagation in determining source parameters. By comparing several events at the same set of stations, empirical station corrections can be made.

Comparing source parameters obtained for the four events so far analyzed, the most apparent result is that the effective stresses and stress drops are almost an order of magnitude larger for event 8 than for the other events. Event 8 is a shallow ($h = 27$ km) located on the main thrust zone between the trench and oceanic front. Events 17 and 20 are intermediately deep earthquakes within the portion of slab of descending Pacific plate landward of the volcanic front, and event 1 is a normal

faulting mechanism located near the trench axis. The high stress drop for event 8 is consistent with the high stress drops of two events analyzed by House and Boatwright (1980), which were also shallow thrust events. One of these previously analyzed events is #9 (Figure 9). The obtained distribution of low- and high-stress drop events is consistent with Archambeau's (1978) stress contour map, in which (static) stress drops of earthquakes were inferred from M_s to m_b ratios. This map (Figure 33) shows an area of high stress earthquakes located south of the western end of the Alaska Peninsula. Lower stress events surround this area on all sides. Further analysis of the events in our remaining data set is necessary in order to see if the high stress events are indeed localized as suggested by Archambeau's map or if they are more a function of location within certain typical features of the subduction zone, e.g. trench, thrust zone, of within descending slab.

In making the calculation of source parameters, one of the more dominant factors is the P velocity at the source which enters into the equation as velocity raised to the power of 5/2. Velocity is strongly depth dependent. Hence, to make accurate estimates of stress drop and moment one requires an accurate determination of depth. In general, the published ISC depths are not reliable. Inspection of numerous short-period records, and obtaining rough estimates of expected amplitudes of pP and sP, one can confidently identify depth phases including water bounces pwP, pwwP, and pWWWP. Depths for these events can then be constrained within a few kilometers.

Another more difficult problem is the determination of rupture velocity which appears as a factor cubed in the equation for effective stress. Following House and Boatwright a rupture velocity of 0.6 times the

Figure 33. Part of stress contour map by Archaibeau (1978) for earthquakes of focal depth 40-55 km. Plotted on the map are the locations and typical values of static stress drop for three events which were analyzed by the process described in the text. H&B is the event studied by House and Boatwright, which has a focal depth of 40 km. Events 1 and 8 have depths of 8 and 17 km, respectively.

Figure 14. Steps in processing seismograms: (a) Plot of digitization of the first 10 seconds of event 17 as recorded at WSSN station ENG; (b) Digitization has been interpolated to equal time intervals and band passed filtered (1.0 to 0.3 Hz); (c) Deconvolved velocity plot. The slope of the line is

$$\left\langle \frac{\dot{u}_c(x,t)}{t} \right\rangle$$

which is used to determine effective stress; (d) Deconvolved displacement plot. Shaded section is the area of displacement pulse, $u_c(x,0)$, which is used to determine the moment.

TABLE 3

Source Parameters for Eastern Aleutian Earthquakes So Far Analyzed

	<u>STATION</u>	<u>EFFECTIVE STRESS</u>	<u>STRESS DROP</u>	<u>MOMENT</u>	<u>FAULT RADIUS</u>
Event 17	COR	50 bars	45 bars	6.0×10^{24} dyne-cm	3.9 km
Depth = 191 km	GOL	139 bars	90 bars	1.6×10^{25} dyne-cm	4.2 km
$M_b = 5.5$	LON	41.4 bars			4.2 km
	QUE	30 bars	21 bars	2.9×10^{24} dyne-cm	3.9 km
	SCP	164 bars		1.5×10^{25} dyne-cm	4.2 km
	TUC	90 bars	77 bars	7.5×10^{24} dyne-cm	3.5 km
Event 24	GSC	26 bars	105 bars	2.15×10^{24} dyne-cm	2.5 km
Depth = 136	GOL	27 bars	41 bars	2.09×10^{24} dyne-cm	2.8 km
$M_b = 5.3$	KTG	131 bars	21 bars	3.67×10^{24} dyne-cm	3.3 km
Event 1	ATL	50 bars			
Depth = 8 km	FVM	116 bars			
$M_b = 5.8$	MDI	100 bars	197 bars	5.9×10^{24} dyne-cm	2.4 km
	PTO	167 bars			2.1 km
	SNG	189 bars			2.4 km
	TOL	99 bars	235 bars	5.13×10^{24} dyne-cm	2.1 km
	UHE	67 bars	138 bars	9.14×10^{24} dyne-cm	2.1 km
	OXF	107 bars			
	WES	129 bars	261 bars	7.8×10^{24} dyne-cm	2.4 km
Event 8	LON	1420 bars			2.1 km
Depth = 17 km	SNG	792 bars			1.4 km
$M_b = 6.1$	STU	1030 bars	948 bars	1.4×10^{25} dyne-cm	2.3 km
	TUC	860 bars	500 bars	1.4×10^{25} dyne-cm	2.3 km
	OXF		728 bars	2.1×10^{25} dyne-cm	2.3 km

Events Analyzed by House and Boatwright (1979)

Depth = 40 km	SDP	1040 bars	840 bars	3.5×10^{24} dyne-cm	1.2 km
$M_b = 5.8$					
Depth = 40 km	SDP	780 bars	640 bars	6.7×10^{24} dyne-cm	1.65 km
$M_b = 6.0$					

shear-wave velocity was used for the shallow thrust event. For the other events a rupture velocity of 0.75 times the shear velocity was used. In order to fix the rupture velocity more accurately one could follow the procedure of matching v^2 plots of synthetics with the data as described by House and Boatwright (1980).

To extend this investigation of source parameters for events of magnitude greater than 6.5, one must turn from using short-period records to long-period records instead. The larger events have the advantage of reducing the station variations which so strongly affect the shorter-period signals. Essentially the analysis of long-period data follows the same procedure as for the short-period records, except the effect of depth phases must be compensated for in the long period pulse shape of the P-wave. Only a few events (see Table 1) will be large enough to be analyzed from long-period records.

Conclusions: This study is aimed to obtain for the Aleutian arc systematically static and dynamic source parameters (e.g. moment, source area, slip, stress drop and ambient stress) for as many earthquakes with magnitudes $m_b \geq 5$ or larger as possible. We believe such data are important to further our understanding of the magnitudes of tectonic stresses in various parts of a subduction zone on the one hand; and—moreover—to categorize strength of seismic sources in the major seismogenic regions of the eastern Aleutian arc to obtain a set of 'typical sources' required in any scheme for seismic risk and hazards analysis.

Crustal Deformation Study

The crustal deformation study is composed of the following elements:

- (1) precise geodetic level lines
- (2) sea-level meter study
- (3) geologic field study of recent faulting and uplifted terraces.

These studies all originated under this DOE study but have obtained since 1978/79 separate funding from NSF or USGS, respectively. Unfortunately support from USGS for the Shumagin Island sea-level meter project is at present uncertain and therefore its future may be severely jeopardized. This is most unfortunate in the light of the seismotectonic findings which confirm the existence of the Shumagin seismic gap and suggest that the occurrence of a large or great earthquake there is significantly high for the next decade.

Leveling results. Several of the level-lines were resurveyed. The most reliable level line, i.e. that with the highest end marker stability, is the Squaw Harbor Level Line on Unga Island. The 1979 results from this line (Figure 35) show a small reversal in the tilting trend which from 1972 to 1978 showed a south-down, north-up tilt at rates of less than half a microradian per year. This reversal is barely significant when considering the scatter of data but if real, it may be of importance as a possible precursory change before a future large earthquake.

Sea-level meter study. The state of sea-level meter study in the Shumagin Islands region as of early 1979 is summarized in a paper published by Bilham and Beavan (1979) which in part is reproduced in the Appendix. Continuation of this project is at present uncertain because of difficulties described above.

Geologic fault and terrace survey. The terrace study and its continuation under NSF funding shows great promise. Figure 36 shows schematically the terraces which were identified on various Shumagin

Figure 36. Schematic display of terrace levels in Shumagin Islands. Observations are arranged as if it were a profile which traverses the islands from trenchward (left) to most landward location on the arc. Far-right terrace profile from the Central Aleutians is taken from the literature for comparison only. Horizontal dimensions of terrace are a qualitative measure for its prominence or persistence on any given island.

Islands during the 1978 field season by M. Winslow. In 1979 an extensive field sampling program on these and some additional terraces were carried out to retrieve datable materials (cores) and count and identify ash layers on them to facilitate correlation of corresponding terraces across islands. The dating methods will include palynology, radiocarbon method (C_{14}) and tephrochronology. More than 40 sites were sampled and the samples have been shipped from the Shumagin Islands to New York for analysis. If the samples yield reliable and consistent dates it will be possible for the first time in the Shumagin Islands to obtain uplift and subsidence rates and to test regional coherence of these vertical motions. If regional coherence can be shown to exist some geologic constraints may be derived for recurrence times of great regional earthquakes in this arc segment.

STATE OF TECHNICAL PROGRAM

Both seismic networks, centered on Sand Point, Shumagin Islands, and Dutch Harbor, Unalaska Island, have undergone considerable technical changes in a basic effort to increase quality, reliability and up-time of the networks, i.e. of their remote stations, telemetry links and central recording facilities. We consider this technical effort still incomplete, although the following changes have been implemented:

Shumagin Network - Central Recording Site. The physical installation at the central recording site in Sand Point was entirely redesigned. A central patch-panel was installed which allows flexibility to record any combination of incoming signals on any combination of recording devices. This flexibility became a necessity for a smooth transition to analog event recording and future digital recording. The analog event recording mode will be implemented during the last quarter of the current contract period.

To avoid further difficulties with timing and clock correction a self-correcting GOES-satellite receiver clock was acquired for this recording site.

Purchase and fabrication of the event-triggered analog tape recording system is completed. Final lab-testing of the unit is presently carried out and installation is planned in December 1979 or January 1980.

Field Stations. A considerable progress has been made to weed out a mixture of old and diverse transmitters and amplifier-VCO's. Because of monetary constraints new Monitron transmitter/receivers could be installed only at those telemetry links which carry multichannel transmission. Associated with this conversion was a switch-over from phase (PM) to frequency (FM) modulation for telemetry. Also newly built VCO mixers were installed. At a total of 12 stations, strong bear-proof, and environmental protection casings were installed to house safely the electronics. To improve radio telemetry stacked antennas were installed at four sites. One station (FPS) was relocated to a more suitable site. At three stations only the seismometer was relocated (IVF, DRR, CNB). At present (November 1979), all stations of the Shumagin, Cold Bay, and Pavlof arrays are believed to be operative except for two of the Pavlof stations (PS3, PN).

Dutch Harbor/Unalaska Network.

Central Recording Site. Installation of the analog event tape recording unit was completed. Due to proximity to strong airport radio transmitters difficulties arose with false triggers of the system from very strong electromagnetic radio interference. Shielding and filtering efforts were only partly successful. Since within the next year the

airport facilities in which the site is located will be torn down we will have to negotiate for a new location of the central recording site which then should avoid radio interference as well. In preparation for this future relocation we have installed a new radio repeater site (BHR) on a high mountain overlooking the entire Dutch Harbor/Unalaska Bay which will allow line-of-site conditions to almost any future recording-site to be selected. This site relays presently the Akutan Volcano station and the newly relocated Makushin Volcano station (formerly MAX, now MKV).

Remaining Tasks. During the last quarter of this contract period, the following technical tasks will be carried out:

- (1) Installation of the analog event-tape recording system at Sand Point;
- (2) Test installation of broad-band channels at Sand Point;
- (3) Construction of an automatic gain-ranging amplifier/VCO demonstration unit.

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APPENDIX I

Papers Published

- (1) Davies, J.N., and L. House, Aleutian subduction zone seismicity, volcano-trench separation, and their relation to great thrust-type earthquakes, Jour. Geophys. Res., 84, 4583-4591, 1979.
- (2) Bilham, R., and J. Beavan, Satellite telemetry of sea-level data to monitor crustal motions in the Shumagin Islands Region of the Aleutian Arc, In: A. Vogel (editor), Terrestrial and Space Techniques in Earthquake Prediction Research; E.S.C.-E.G.S. Special, Conference Proceedings, F. Vieweg and Sons, Braunschweig/Wiesbaden, 1979, pages 269-283.

Papers Submitted for Publication

- (1) McCann, W.R., O.J. Perez, and L.R. Sykes, Yakataga Seismic Gap, southern Alaska: History and earthquake potential, Science, in press, 1979.
- (2) Perez, O.J., and K.H. Jacob, Tectonic model and seismic potential of the eastern Gulf of Alaska and Yakataga Seismic Gap, submitted to Jour. Geophys. Res., October, 1979.
- (3) House, L., and J. Boatwright, Investigation of two high stress-drop earthquakes in the Shumagin Seismic Gap, Alaska, submitted to Jour. Geophys. Res., 1979.
- (4) Boatwright, J., Preliminary body wave analysis of the St. Elias, Alaska earthquake of February 28, 1979, submitted to Bull. Seismol. Soc. Amer., 1979.

Papers in Preparation

- (1) Davies, J.N., L.R. Sykes, L. House, and K.H. Jacob, Evidence of a seismic gap in the Shumagin Islands region of the eastern Aleutian arc, Alaska, to be submitted to Jour. Geophys. Res., 1980.
- (2) House, L., and K.H. Jacob, Earthquake focal mechanisms and tectonics of the eastern Aleutian arc, in preparation, 1980.
- (3) Perez, O.J., and K.H. Jacob, St. Elias, Alaska, earthquake of February 28, 1979: Tectonic setting and seismic pattern prior to it, to be submitted to Bull. Seismol. Soc. Amer., 1980.

SEISMOTECTONIC SETTING OF THE SHUMAGIN ISLANDS
REGION, ALASKA: EVIDENCE FOR A SEISMIC GAP AND A
REGION OF HIGH SEISMIC POTENTIAL

J.N. Davies (Lamont-Doherty Geological Observa-
tory, Columbia University, Palisades, N.Y.
10964)

L. House (Lamont-Doherty Geological Observatory
and Dept. of Geol. Sciences, Columbia Univer-
sity, Palisades, N.Y. 10964)

The Shumagin Islands are approximately in the center of a geographical area of the Aleutian arc that extends from Unalaska of the eastern Fox Islands to the western end of Kodiak Island. We present a study of the seismotectonic setting of this area that focuses on the Shumagin Islands region in which we have operated a local seismograph network since 1973. This area is bounded by the aftershock zones of the great earthquake of 1957 in the Andreanof-Fox Islands to the west and that of the 1964 in the Kodiak Island-Prince William Sound region to the east. The eastern half of this Unalaska-to-Kodiak area broke in the great earthquake of 1938. The tsunami earthquake of 1946 occurred in the western part of the area but most likely did not break much, if any, of the interplate boundary. It is along the interplate boundary above 50 km depth, i.e. along the main-thrust zone that great, thrust-type earthquakes occur. New maps of FDE and ISC epicenters for the eastern Aleutian arc suggest that within the past decade a "Mogi-Donut" pattern has become fairly well established around the Shumagin Islands region. The down-dip part of the donut pattern (at 40 km depth) is best developed and consists, at least in part, of events that have a very high stress drops (order of 1 kilobar) and which have focal mechanisms indicating underthrusting along a plane dipping at 30°. On the basis of these combined data, both historic and recent, we conclude that a seismic gap exists in the Shumagin Islands region which has a relatively high potential for an earthquake of magnitude ≥ 8 within the next few decades.

1. 1979 Fall Meeting AGU
2. 023052DAVIES
3. J.N. Davies
Lamont-Doherty Geo. Obs.
Palisades, N.Y. 10964
4. S
5. None
6. None
7. 40%
8. Mrs. Alma Kasner
Purchasing Dept.
Lamont-Doherty Geo. Obs.
Palisades, N.Y. 10964
9. C

VOLCANIC EARTHQUAKES CORRELATED WITH EARTH TIDES AT PAVLOF VOLCANO

S. McNutt (Dept. Geol. Sci. and Lamont-Doherty
Geological Observatory of Columbia University,
Palisades, New York 10964)
(Sponsor: Klaus Jacob)

A single short-period seismometer has been operated 7 km to the southeast of the summit of Pavlof Volcano since October 1973. Continuous helicorder records from an eruption sequence beginning 29 October 1974 and ending 13 January 1975 show four types of seismic signals: micro-earthquakes (A-type), volcanic tremors (B-type), explosion quakes with distinct air phases, and harmonic tremors. Histograms of the number of events of each type that took place during each two-hour interval are compared with the horizontal component of the theoretical solid-earth tidal strain, calculated for an azimuth of 135° (subparallel to the direction of plate convergence). For 3-4 days immediately before and after swarms of explosion quakes, a strong positive correlation is observed between B-type earthquake swarms and semi-diurnal tidal peaks. Before the explosions occur, swarms of B-type quakes correlate with the maximum tensional (high tide) peak, and after the explosions stop, swarms correlate with the maximum compressional (low tide) peak. This information strongly suggests that B-type earthquakes are caused by expansion and contraction, respectively, of shallow-depth magma conduits, and demonstrates that just before and just after eruptions the volcano is very sensitive and responsive to small changes in ambient stresses.

1. 021138JACOB (sponsor)
2. 1979 NW Section Meeting U
Bend, Oregon
3. Volcanology
4. Volcanology
5. No
6. No
7. None
8. Bill to:
Alma Kesner, Purchasing
Lamont-Doherty Geol. Obs.
Palisades, NY 10964

TELESEISMIC P-RESIDUALS AND RAY TRACING IN THE
ALEUTIANS AND SOUTHERN ALASKA: EVIDENCE FOR
LATERAL INHOMOGENEITIES IN THE CRUST AND UPPER
MANTLE

J. Krause (Lamont-Doherty Geological Observatory
of Columbia University and Dept. of Geol.
Sciences, Palisades, N.Y. 10964)

K. Jacob (Lamont-Doherty Geological Observatory
of Columbia University, Palisades, N.Y. 10964)

P-wave residuals with respect to a Harrin (1968) earth model for regional and teleseismic events recorded at stations in the Aleutians and southern Alaska show strong dependence on azimuth. For explosions on Amchitka Island, residuals are negative (i.e. arrivals are early) by up to 6 sec. Residuals from earthquakes in the Shumagin Islands region are generally positive by several seconds. Residuals from Soviet explosions on Novaya Zemlya range from negative by ≤ 1 sec. in the Aleutian arc to positive by 2-3 sec. in the Gulf of Alaska. Relative residuals for this source region with respect to station GMA in western Alaska are increasingly negative towards the west in the Aleutian arc. Residuals for the San Fernando, California, earthquake are positive by up to 5 sec.

Many of these variations in P-residuals can be explained by the high-velocity slab of lithosphere dipping beneath the Aleutians and southern Alaska. Rays which travel a large distance laterally along the dipping slab produce large negative residuals; those travelling up-dip along the slab produce small negative residuals; rays which miss the slab entirely appear to be influenced by other, weaker inhomogeneities in the upper mantle, and produce generally positive residuals.

We trace rays through a slab model derived from hypocenter locations to arrive at a velocity structure which approximately fits the observed travel times.

1. 1979 Fall Meeting AGU
2. 039516KRAUSE
3. Janet Krause
Lamont-Doherty Geo. Obs.
Palisades, New York 10964
4. S
- 5.
6. 0
7. 0%
8. Mrs. Alma Kesner
Purchasing Department
Lamont-Doherty Geo. Obs.
Palisades, New York 10964

student rate applicable

SEISMIC HISTORY AND EARTHQUAKE POTENTIAL OF YAKATAGA SEISMIC GAP, SOUTHERN ALASKA

W. R. McCann

O. J. Pérez

L. R. Sykes (all at: Lamont-Doherty Geological
Observatory and Department of Geological
Sciences of Columbia University, Palisades,
New York 10964)

Several authors have identified segments of the North American-Pacific plate boundary in southern Alaska and the Aleutians as seismic gaps. One gap is bounded by the rupture zones of the great earthquakes of 1958 and 1964. Only a small part of this gap, however, broke during the large ($M_s \sim 7.7$) St. Elias earthquake of 28 February 1979. A re-evaluation of the seismic intensity data indicates that about 250 km of the plate boundary broke during the series of great events on September 4 and 10, 1899. Focal mechanisms of earthquakes in this region show shallow thrusting in a northerly direction. These mechanisms and geologic data indicate a décollement-style deformation with a series of imbricate thrusts in the upper plate. Moderate-size shocks since 1964 define a ring of activity surrounding a zone of near quiescence. This ring of activity is spatially similar to the "Mogi-donut" pattern observed before several great earthquakes. The average slip calculated for the 1899 sequence is comparable to the potential slip built up since then by plate movements. Although none of these methods yields a precise estimate of the time of occurrence of the next great earthquake in the Yakataga gap, the region merits intensive study for possible precursory effects.

1. 033692MC CANN
2. 1979 Spring AGU
3. Seismology
4. None
5. No
6. No
7. 0%
8. Bill to:
Alma Kesner, Purchasing
Lamont-Doherty Geol. Obs.
Palisades, NY 10964

THE TECTONIC SETTING OF THE LARGE GULF OF
ALASKA EARTHQUAKE OF FEBRUARY 28, 1979

Omar Perez (also at: Dept. Geological Sci.,
Columbia University)
Klaus H. Jacob (both at: Lamont-Doherty Geol.
Obs., Columbia U., Palisades, NY 10964)

On February 28, 1979 at about 21h 27m a large earthquake (M_s near to 8.0) occurred in the Gulf of Alaska coastal region. The occurrence of this large earthquake was anticipated on the basis of the seismic gap theory (Sykes, 1971; Page, 1971; Kelleher *et al.*, 1973). A study of all earthquakes during the period 1964-1975 which yielded fault-plane solutions using teleseismic information was completed just prior to this large event. All earthquakes indicated shallow-angle thrusting of the Pacific plate in a northerly direction both beneath the offshore Tertiary province (along a décollement zone) and on the onshore regions of the North American plate beneath the Chugach-St. Elias Ranges. From geologic evidence and from surface faulting related to the Sept. 10, 1899 earthquake in Yakutat Bay, it is known that steep-angle reverse faulting, associated with crustal-shortening and uplift of the northern blocks, can also occur in this region. Based on these interpretations we concur with McCann *et al.* (1979) in assigning a high seismic potential to this plate boundary segment, but we suggest that the area can be in general the site of two different types of earthquakes: shallow thrusting of Pacific Ocean floor in the offshore region and on-land, and steep-angle reverse faulting associated with the collisional boundary inland. It therefore seems possible that other large earthquakes following the February 1979 event may still occur in the region, similar to the series of great events that took place in the area around the turn of the century. In this interpretation the multiple events are related to successive rupture of the two stress-relieving features (décollement and boundary fault, respectively) at this plate margin.

1. 702418PEREZ
2. 1979 Spring Meeting AGU
3. Seismology
4. None
5. No
6. No
7. 0%
8. Bill to:
Alma Kesner, Purchasing
Lamont-Doherty Geol. Obs.
Palisades, NY 10964

PRELIMINARY BODY WAVE ANALYSIS OF THE ST.
ELIAS, ALASKA, EARTHQUAKE OF FEBRUARY 28, 1978

John Boatwright

Omar Perez (both at: Lamont-Doherty Geological
Observatory and Department of Geological
Sciences of Columbia University, Palisades,
New York 10964)

We have made a preliminary analysis of the body waves radiated by the St. Elias earthquake, using long period P- and S-waves recorded at three WWSSN stations and at Palisades, N.Y. The focal mechanism, determined from P-wave first motions and S-wave polarizations, shows a very shallow angle thrust striking 70°E and dipping 10°W . Using this mechanism and a best average source depth of ≈ 11 km, we have directly deconvolved the interference of the depth phases (i.e., pP and sP, or sS) from the recorded pulse shapes, thus obtaining approximate whole space (AWS) velocity and displacement pulse shapes. These AWS pulse shapes show that the earthquake contained three distinct rupture events as well as a small initial event ≈ 7 secs earlier. The earthquake propagated to the southeast. From the pulse rise times of these sub-events, the rupture lengths for these events are $\approx 9, 24$ and 16 km, assuming that the rupture velocity was 2.5 km/sec. The cumulative moment for the whole event obtained from this analysis, 1.2×10^{27} dyne-cm, is substantially smaller than the surface wave moments obtained by Hasegawa, and Stevens and Boore, of $\approx 5 \times 10^{27}$ dyne-cm (personal communications). The moments of sub-events are estimated to be $\approx .8, 3.5$ and 7.6×10^{26} dyne-cm, respectively.

1. 702418PEREZ
2. 1979 Spring Meeting AGU
3. Seismology
4. Special Late Friday Session
5. No
6. No
7. None
8. (Mrs.) Alma Kesner
Purchasing Dept.,
Lamont-Doherty Geol. Obs.
Palisades, NY 10964

SEISMOTECTONIC SETTING OF THE SHUMAGIN ISLANDS REGION, ALASKA:
EVIDENCE FOR A SEISMIC GAP AND A REGION OF HIGH SEISMIC
POTENTIAL*

John Davies, Lynn Sykes, Leigh House
and Klaus Jacob
LamontDoherty Geological Observatory
of Columbia University
Palisades, New York 10964

The Shumagin Islands are approximately in the center of a geographical area of the Aleutian arc that extends from Unalaska of the eastern Fox Islands to the western end of Kodiak Island. We present a study of the seismotectonic setting of this area that focuses on the Shumagin Islands region in which we have operated a local seismograph network since 1973. This area is bounded by the aftershock zones of the great earthquake of 1957 in the Andreanof - Fox Islands to the west and that of 1964 in the Kodiak Island - Prince William Sound region to the east. The eastern half of this Unalaska-to-Kodiak area broke in the great earthquake of 1938. The tsunami earthquake of 1946 occurred in the western part of the area but most likely did not break much, if any, of the interplate boundary. It is along the interplate boundary above 50 km depth, i.e., along the main-thrust zone that great, thrust-type earthquakes occur. We show that the down-dip width of the main-thrust zone is proportional to the recurrence times for great thrust-type earthquakes along the Alaska-Aleutian convergence zone. In the Shumagin Islands region the width of the main-thrust zone is intermediate to those in the adjacent regions. This implies recurrence intervals on the order of 80 to 140 years as compared to 50 to 80 years for the central and western Aleutians and about 220 years for the Kenai Peninsula - Prince William Sound region. These estimates of recurrence intervals are based in part on new historic seismicity data and in part on rates of plate convergence. New maps of PDE and ISC epicenters for the eastern Aleutian arc suggest that within the past decade a "Mogi-Donut" pattern has become fairly well established around the Shumagin Islands region. The down-dip part of the donut pattern (at about 40 km depth) is best developed and consists, at least in part, of events that indicate a very high stress drop (order of 1 kilobar) and which have focal mechanisms indicating underthrusting along a plane dipping at 30°. On the basis of these combined data, both historic and recent, we conclude that a seismic gap exists in the Shumagin Islands region which has a relatively high potential for an earthquake of magnitude greater or equal to 8 within the next few decades. A questionable gap in the eastern Fox Islands region may have not broken in the great shock of 1957; it may or may not be connected with the Shumagin Gap.

*Proceedings 30th Alaska Science Conference, September 19-21, 1979, Alaska

A TECTONIC MODEL FOR THE EASTERN GULF OF ALASKA AND ITS IMPLICATIONS FOR SEISMIC HAZARDS*

Klaus H. Jacob and Omar J. Perez
Lamont-Doherty Geological Observatory
of Columbia University
Palisades, New York 10964

A tectonic model is derived for the motions and deformations along the Pacific North American plate boundary in the Eastern Gulf of Alaska. The major data base consists of 13 new and several published fault-plane solutions, and of published regional plate-motion vectors obtained from global inversions.

The model predicts a component of about 1.3 cm/year of convergence across this predominant-transform boundary (associated with 5.3 to 5.7 cm/year of right-lateral strike-slip motion along the Queen Charlotte Fairweather fault systems). This small convergence leads, however, to a slow subduction of Pacific lithosphere beneath this western North American margin. A dipping Benioff zone is absent there but a weakly developed volcanic arc exists in the northern third of the boundary where the lithosphere is expected to extend deeper than 100 km into the mantle. West of Yakutat Bay convergence is dominant and is taken up along several imbricate thrust belts. Probably 80% of the 6 cm/year convergence are taken up on shallow-angle thrusts in the Chugach St. Elias fault system extending from the Pamplona zone offshore to the Border Ranges fault inland. A segment of this plate boundary has been identified as the "Yakataga seismic gap". It has not ruptured by a large earthquake for 80 years and is expected to rupture in a large quake in the not too distant future. Recurrence times at this plate-boundary segment are estimated to be near 100 years (with uncertainties ranging from 50 to 200 years). The St. Elias earthquake of February 28, 1979, ($M_s = 7.7$) relieved stresses over an area much smaller than that of the Yakataga seismic gap.

A secondary shallow-angle thrust zone exists offshore along the northwest-trending shelf-edge connecting the Queen Charlotte fault (near Cross Sound) to the Aleutian trench (near Kayak Island). Slip is northeast-directed with rates estimated to be about 1.2 cm/year. Recurrence times are estimated to be 500 years but are highly uncertain.

Recurrence times for presently very weakly active faults inland (i.e., Denali fault, Totschunda Fault amongst others) may be even longer (approximately 1,000 years?).

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Satellite Telemetry of Sea-Level Data to Monitor Crustal Motions in the Shumagin Islands Region of the Aleutian Arc

Roger Gilliam and John Seaward

Lamont-Daherty Geological Observatory of Columbia University, Palisades, New York 10964 USA

SUMMARY

Sea level is being used to provide a datum for monitoring vertical tectonic deformation in the Shumagin Islands Seismic Gap. An array of sea-level monitors has been installed that records filtered near-surface sea-level variations and in some cases, piezometric pressures on shore-lines with a resolution of 1 mm of water height. The data from the sea-level monitors are transmitted via satellite to provide hourly values for analysis. The spacing between monitors is less than 30 km to enable deformation processes to be studied in detail. In the past two years, the Islands have tilted to the south with an amplitude of approximately 10^{-5} radians/year.

INTRODUCTION

In this article we describe experimental measurements of mean sea-level to monitor vertical tectonic motions in the Aleutian Shumagin Islands, a seismic gap in the eastern Aleutian Islands arc, Alaska.

Seismic gap theory has proved to be remarkably successful at predicting the approximate location and timing of future great earthquakes (Kelliker [1], Kelliker *et al.* [2], McGinn *et al.* [3]). The method involves examining the spatial or temporal pattern of energy release historically exhibited by a seismic belt in order to identify regions that appear to be overdue for a major event. The precise location and time of such an event cannot be predicted without additional information describing physical conditions in the source region. A major

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source of data indicating a relationship between seismicity and crustal deformation has traditionally been the measurement of sea level. Crustal deformation is dramatically illustrated in many parts of the world by raised beaches and submerged cities and forests. There are also examples of pre-seismic and post-seismic movements (Omori [4], Inamura [5], Yanaguti [6], Fitch and Scholz [7], Mezherikov [8], Myer [9,10], Scholz and Koro [11], Brown *et al.* [12]).

There are two shortcomings to sea-level data when used to study crustal deformation; the distribution of tide gauges is frequently poor and insufficiently dense to provide spatial details of the deformation process, and sea-level has numerous sources of noise that render it far from being the stable geopotential surface that one would like for precise measurements.

location of sea-level measurements

The present distribution of tide gauges is determined by the oceanographer rather than the seismologist. These gauges are placed in locations selected to give useful tidal data over a wide region. Frequently instruments are spaced several hundred km apart or farther if no suitable harbor exists. In plate tectonic studies the minimum desirable density of sea-level monitors is related both to the size of the earthquake and its mechanism. In the present study we are concerned with events greater than $M_0 = 7$ (Kanamori [13]) and with rupture lengths greater than 50 km. We assume that surface deformation will be detectable out to several fault lengths and estimate that an instrument spacing of between 10 and 30 km may provide satisfactory details of the deformation process. Thus, in comparison to oceanographic requirements, a denser distribution of instruments is called for and ideally one that is uniformly spaced on the two-dimensional surface overlaying the epicenter. This is rarely possible and even with many more tide gauges most coastlines will represent only a fraction of the surface of interest. Improvements to absolute pressure transducers may enable measurements of sea level in deep-water regions if monitoring and maintenance difficulties can be overcome, but it seems that, at present, conventional gauges offer a practical method in suitably selected areas of interest.

Archipelagos, island arcs and regions with indented or irregular coastlines appear most promising as areas for immediate study. One such region is the Shumagin Islands in the Aleutian arc (Fig. 1). In this region a chain of islands stretches more than 100 km toward the

Figure 1. Location of tide gauges and mean sea-level monitors in the Gulf of Alaska and the Shumagin Islands. The region to the west of 161°W experienced a major earthquake in 1957 and the region between 155°W and 165°W broke in 1964. A major earthquake may have partly released strain energy in the eastern Shumagin Islands in 1918 (McCann *et al.* [1]).

trench from the mainland. The most southerly land mass is approximately 120 km from the deepest point of the trench and seismic studies have shown the Benioff zone to be less than 30 km deep at this southerly point, deepening rapidly northward (Davies and Howe [14]). The region is identified as a subduction zone and is sited between the 1957 Andreanof-Fox Island earthquake and the 1964 Alaska earthquake, both with magnitudes greater than 8. An earthquake in 1938 may have broken into part of the Shumagin region.

Noise

In tectonic studies any departure of the sea-level surface from the geoid is considered noise. Thus tides are as much noise as are the effects of storm surges, salinity variations, sea-current effects, atmospheric pressure changes and many more variables that can disturb the sea surface. Relatively few of these noise sources can be removed numerically by measuring the appropriate variable (e.g., atmospheric pressure) and empirical methods designed to compensate for noise can only be partially successful since most of the processes are non-stationary. Fortunately, in a limited area sea-level noise is reasonably uniform. For example, Ingraham *et al.* [15] present analyses of data from the Gulf of Alaska indicating good coherence in sea-level over a wide region (Fig. 2). The annual variation of sea level has a range of 20 ± 5 cm with a minimum in May at six gauges spaced on average 300 km apart. Since we are interested in differences in sea level, i.e., variations in a region less than 200 km in extent, we hope to reduce much of the coherent noise present. We hope to reduce some sources of incoherent noise also by installing more than one instrument on an island. Where possible the sea-level monitors face the sea in opposing directions such that a mean value partially eliminates noise introduced by changes in current direction. A future possibility is that space borne laser altimetry will provide an independent definition of the mean sea-surface. The present accuracy of satellite altimetry is insufficient to be useful (± 10 cm) since present sea-level analysis methods appear to attain 1 cm accuracy even when widely-spaced tide gauges are used (Myer [9]). Frequent vertical deformation amounting to tens of cm has been reported for some earthquakes (Scholz and Katz [11], Myer [10]) although there are many examples where any deformation that has occurred is below the noise level of affected tide-gauges. For example, whereas post-seismic deformation is evidently associated with the 1957 Fox

Figure 2. Annual mean sea-level data (1950-1974) from six National Ocean Survey Gauges in the Gulf of Alaska. The data have been smoothed using a 82 month running mean (Ingraham *et al.* [15]). The Unalaska data after 1957 and the Kodiak data after 1964 show post-seismic land emergence following great earthquakes ($M_s > 8$). The Kodiak and Seward data from 1964 on reflect post-seismic movements that appear also on geodetic levelling data (Brown *et al.* [12]).

The lower figure shows the coherence for monthly mean sea-level data between Ketchikan and Unalaska, two widely separated gauges (700 km) and between Ketchikan and Sitka, the closest two gauges (320 km). Coherence clearly falls off with distance though even at 700 km coherence is shown at annual, semi-annual and tri-annual periods.

Island and 1964 Alaska earthquakes (Fig. 2) there is no obvious precursory sea-level change. Thus the precision desired of a sea-level monitor appears to be of the order of 1 mm. It remains to be seen whether noise sources even in a limited region can be suppressed to this level.

MEAN SEA-LEVEL MONITOR

The sea-level monitor is described by Bilham [16]. The device can be installed on remote coastlines and does not require a jetty or harbor. The system operates from batteries or solar-power and requires a minimum of maintenance. It differs from a conventional tide-gauge in that the tides are filtered prior to recording and that it has a resolution of 1 mm in sea-level change. We discuss various problems we have encountered in the operation of the gauge and corresponding improvements to the instrument.

The instrument consists of a vertical tube of about 15 cm diameter buried inland and connected to the sea via a long thin entry tube and a hydraulic filter (Fig. 3). Sea-level variations are monitored in the vertical tube using a float and pulley arrangement. The hydraulic time constant of the gauge is chosen to be large enough to suppress the amplitude of diurnal and semi-diurnal tides to less than 10 cm. Filtering was originally achieved using a length of capillary tubing such that the time constant was given by $\tau = 8\eta\ell^2/r^4g$ seconds where ℓ is the length of capillary tube of radius r cm connected to a standpipe of radius ℓ cm, η is the kinematic viscosity of sea water (sea water = $.01 \text{ cm}^2\text{s}^{-1}$) and g the acceleration due to gravity. Time constants of the order of 80 hours were found convenient to reduce diurnal tidal amplitudes to approximately 5 cm without attenuating monthly sea-level changes by more than 2%. An alternative method of filtering was to use a porous filter of large area. The 5 cm diameter micropore disc is not clogged by particles of grit in the sea entry tube, which was a problem with the capillary. The micropore filter, typically with a pore size of 0.2 to 0.5 μm , is also sturdier than the capillary but suffers from the disadvantages that it slowly becomes clogged with finely suspended particles so that its time constant increases with time. Fig. 4 shows the percentage amplitude reduction at various periods of sea-level change.

A fundamental problem with the entry tube is that air envelopes from the highly aerated zoogeal waters and migrated as bubbles to the highest

Figure 3. (a) Mean sea-level monitor showing hydraulic filtering in entry tube to sea. The entry tube also incorporates a bubble trap to prolong the life of the system in the presence of aerated coastal water. (b) Piezometric monitor located on beach. A coarse filter prevents ingress of mud and sand. Variations in the piezometric surface are related to mean sea-level variations, although inland hydrology is a major noise source on many coasts.

Figure 4. Attenuation of sea level amplitude as a function of period introduced by hydraulic filtering with an 80 hour time constant.

point of the syphon, eventually blocking the tube. Repeated visits were found necessary to evacuate this air and a bubble-trap was devised to extend the useful life of the syphon. This consisted of a large container in the hydraulic line positioned above the remotest point of the syphon (Fig. 7). The tube remains full of water in this arrangement until the bubble trap empties. Visits at four-month intervals to refill the bubble trap appear to be necessary to maintain syphon operation. The evacuation of bubbles is not a problem if the entry tube can be arranged to incline gently downwards from the base of the steepside. This may be possible with the assistance of earth moving machinery but in remote island locations where the installations are invariably by hand it is seldom possible to avoid a syphon arrangement.

An alternative method of avoiding gaseous exsolution in the entry tube is to abandon the direct connection with the sea entirely. This has been tried experimentally in several locations where we monitor water level in a drilling well on the beach. Theoretically, the sand water level in a uniform porous coastal aquifer defines a mean piezometric surface that is equivalent to mean sea level. In practice there are few locations where the piezometric surface is not dominated by in-

land hydrology. Such is the simplicity of piezometric measurements, however, that we considered it worthwhile to conduct simultaneous measurements of sea level and inland water level to determine their long-term relationship.

Tidal variations in coastal aquifers are caused by pressure variations at the land-sea interface (Van der Kamp [17]). Only at the boundary and a short distance inland does any flow occur. The higher the porosity the slower the decay in tidal variation inland (Fig. 5). During many of our installations we encountered the water table on the beach at approximately mean sea level only metres from the surf. The stability of the water table in some sites is indicated by a marked change in the soil chemistry, from an oxidizing to a reducing environment, a change that occurs over a few cm near the water table (e.g., Pirate Cove, Papoi Island). In other areas of higher porosity, e.g., on a gravel spit at Eagle Harbor, Nagai Island the tide was so well-coupled that the underground water-table dropped well below mean sea-level on the outgoing tide.

Figure 5. Phase and amplitude relationship observed at various distances inland on a porous sand deposit, West End, Anegada.

The mean piezometric water-level inland in a porous aquifer is higher than sea level primarily due to rainfall raising the piezometric surface. The elevation is covered by a "steady-state" fresh water lens in equilibrium with the surrounding higher density sea water and secondly by a transient additional amount of water resulting from recent rain. Both effects can be reduced by choosing a small landmass to undertake measurements (Fig. 6). Alternatively spite of land, narrow headlands, and sand bars have the necessary properties unless they are impounding fresh water (e.g., storm beaches) or near river estuaries. The maximum elevation of the piezometric surface by fresh water is approximately 0.05 of the depression of the saline boundary. If we assume that the fresh water lens is unstable in the presence of sea currents if its depth is more than one third its width, then the fresh water elevation error can be estimated to be approximately 3 cm at the center of a 30 m wide spit. If measurements are conducted near the edge of the spit and an allowance is made for partial contamination of the fresh water, the static error reduces to less than 1 cm. The stability of the lens of fresh water over periods of several weeks and longer is of fundamental

Figure 6. Upper figure: hydrological conditions at a porous boundary in the presence of tides. Flow occurs only in the region near the sea. Inland effects are principally fluid pressure changes. Lower figure: the fresh water lens beneath a small island.

importance if the measurements are to reflect sea-level variations, and since this will be unknown at most locations it appears advisable to locate piezometric gauges close to the beach where variations are less significant.

DATA RECORDING AND SATELLITE TRANSMISSION

Sea-level data are recorded locally along with air temperature on a strip chart recorder driven by a quartz clock. The angular-position transducer that records the standpipe fluid-level also provides a 0-5V output that is taken to a LaBarge data collection platform (DCF). The signal is stored in a memory that updates every few hours but which is transmitted in its entirety every 30s using an omnidirectional 600 MHz antenna. At times when the Landsat 1 satellite can see both the DCF and either the Gilmore or Goldstone tracking station, data are relayed to the Goddard Space Flight Center, Maryland. Once a week the data from the entire array of sea-level monitors are forwarded by mail to Lamont-Doherty for analysis.

A sample of satellite data is shown in Fig. 7. Although the Landsat is in polar orbit and therefore can "see" the Shumagin almost every hour, on some passes the sea-level monitors are obscured by mountainous terrain near the site. The lost data are not important and could be reduced either by reusing the antenna position or by selecting sites with lower horizons. Alternatively the DCF's can be programmed to transmit to the geostationary GOES satellite in locations with a southerly horizon.

The DCF operates on solar power and can send eight, eight-bit words per transmission in addition to station identification information. Typically the Landsat satellite relays two or three transmissions during its traverse from horizon to horizon, hence it is possible to double the information transmitted from each site. Routinely we transmit the battery voltage level in addition to sea level and/or piezometric level from each site. The benefits of satellite transmission are twofold: we are able to receive data from a remote network of islands with a minimal delay and we are able to monitor instrument malfunction. The second of these has proved to be invaluable in planning field maintenance trips. The Shumagin Islands are difficult to visit due to the exceptionally bad weather common to this part of Alaska. Many sites can be visited routinely only once per year and at great expense. A malfunction

Figure 7. Sea-level data received via satellite from four locations in the Shumagin Islands between July and December 1978. The instruments have hydraulic time-constants of approximately 200 hours. The lower two records are from piezometric monitors. On the lowest record the gauge was malfunctioning from soon after installation until day 322 when it was visited. The vertical scale on these plots is 1000 units equals 40 cm.

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In a gauge can result in a data gap of many months and may jeopardize the entire experiment. The weekly data provide an immediate check on the number of gauges operating and corrective measures can be initiated when appropriate.

The delay in the data reception caused by mailing time is not critical and can be bypassed if more rapid data transmission is required. We have not considered this to be necessary since the sea-level monitors have time-constants of approximately 3 days and the tectonic precursors we seek extend over months.

DATA

In 1976 we installed two preliminary gauges. These were incorporated into a network of five in 1977 and in 1978 satellite transmission and piezometric measurements were added. Atmospheric pressure is monitored on Popof Island and Sismomof Island. In Fig. 8 we show data obtained in 1977 from these two islands. The two gauges on Popof Island track each other reasonably well but the gauge on Sismomof diverges and apparently indicates a tilt to the south of several micro radians. This is consistent with levelling measurements made on two level lines near Popof Island and is typical of the downwarping shown at underthrust plate margins prior to seismic slip (Scholz and Kato [10]). In Fig. 7 we present data collected since the satellite relay was initiated.

Figure 8

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CONCLUSIONS

The measurement of hydraulically-filtered sea-level variations is potentially capable of monitoring vertical tectonic motions in island arcs. While we have insufficient data presently to determine the accuracy of our measurements, it is probable that monthly mean values will be accurate to within 5 mm. This corresponds to better than 10^{-7} radians tilt precision across the Shumagin Islands. Long-term comparisons between shoreline piezometric measurements and sea level are being conducted to determine if simpler instruments may be possible for future work. The use of satellites to relay data for immediate analysis and to check instrument performance makes the operation of a remote network more reliable than was hitherto possible. During the next decade a major earthquake may occur in or near the Shumagin Islands. The ability to monitor surface deformation continuously and remotely through the pre-seismic and post-seismic stages is an exciting advance in tectonic studies.

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APPENDIX IVTime Devoted By Principal InvestigatorsFirst three quarters (March 1 to November 30, 1979):

J.N. Davies	9.0 months	(100% of time)
K.H. Jacob	5.6 months	(70% of time)
L.R. Sykes	0.5 months	(10% of time)

Expected time devoted during final quarter:

J.N. Davies	0 months	(0% of time)
K.H. Jacob	.4 months	(13% of time)
L.R. Sykes	.5 months	(17% of time)